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THE INTERNAL STRUCTURE OF THE EARTH, MOON, AND PLANETS

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16. Abstract

In this brochure geophysical methods are discussed which have been applied to the study of the internal structure of the earth. Seismology, gravimetry, the geomagnetic field, geothermics, and the study of substances at high pressures are also discussed. A description is given of proposed models of the internal structure of the earth and of the other planets in the solar system. The brochure is intended for specialists in the field, geophysical engineers, students, and all those who are interested in contemporary problems of geophysics.

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THE INTERNAL STRUCTURE OF THE EARTH, MOON, AND PLANETS

by V. N. Zharkov (Doctor of Physical-Mathematical Sciences)

Table of Contents

". . . Sciences which are not derived from experiment, this foundation of all knowledge, are useless and full of errors . . ." Leonardo da Vinci

INTRODUCTION

The study of the internal structure of the earth and the /3*
planets and their evolution is one of the central problems of modern science. The water and air coverings of our planet, as well as the solid earth crust, are secondary products of the development of the earth--they all have been produced from the interior of our planet during the course of its geological history. Therefore, in order to understand better how these three external coverings of the earth have been built up, scientists study the structure of the earth's interior. The whole problem, roughly speaking, is divided into two parts. The first and simpler part is just the question of how the interior of the earth is structured in the present time; the second and more complicated part is the question of how the earth was structured previously and what changes it has undergone during the time of its existence, equal to some $4.5 \cdot 10^6$ years.

The most important feature of geophysics--the science that uses physical methods to study the earth--is the fact that, of necessity, a large amount of work is based on theoretical methods, inasmuch as direct penetration into the interior of the earth is not possible.

Of course, one might think of geophysics as a purely theoretical science. No . . . geophysics as a branch of natural science is based on experimental geophysical data, but all these data are always indirect. Only theoretical analysis of geophysical data makes it possible for us to make judgments about this or that property of the earth's interior.

Geophysical research itself is much more complicated than pure physical research. The fact is that the physicist in his laboratory sets up an experiment so that it is convenient for him to study this or that aspect of some phenomenon under consideration. The geophysicist is deprived of such a "luxury." For him, nature herself, for the most part, sets up the experiments.
Thus, the geophysicist must await events (earthquakes or electro- /4
magnetic storms) in order to study the signal that arises from the phenomenon under investigation and in order to study the results of the passage of such signals through the planet so as to be able, after analyzing them, to make the necessary conclusions. These circumstances explain the complexity of geophysical experimentation.

*Numbers in the margin indicate pagination in the foreign text.

A natural question arises: just why in geophysics would one not set up the same sort of planned experiments as is the accepted practice in physics? Actually, we do use artificial explosions as a source of seismic waves and we do place them in regions where we wish to carry out our work and we do locate our recording devices in the most convenient fashion. Thus, these artificial explosions are used in the seismic search for useful minerals, in particular, in the search for petroleum. However, the fact still remains that the energy given off in an earthquake is a thousand times higher than the energy of an artificial explosion and this high energy makes it possible to probe the earth as a whole. We refrain here from discussion of large nuclear explosions, which still cannot be compared in magnitude with earthquakes.

Therefore, scientists strive to use all possibilities to get information on the earth's interior. Geophysical research is always complex, that is, it is carried on by using a variety of methods. We attempt in this brochure to discuss the various aspects of complex geophysical research. The reader who wishes a more detailed knowledge of these questions is referred to the following books: V. A. Magnitskiy, "Internal Structure and the Physics of the Earth" ("Nedra" Publishing House, Moscow, 1965); V. N. Zharkov, V. P. Trubitsyn, and L. V. Samsonenko, "Physics of the Earth and the Planets. Shapes and Internal Structure" ("Nauka" Publishing House, Moscow, 1971).

GEOPHYSICAL METHODS FOR STUDYING THE EARTH'S INTERIOR

Seismology. The Seismic Model of the Earth

For a long time, seismology, one of whose founders was the Russian physicist and geophysicist academician Boris Borisovich Golitsyn, was the science of earthquakes and seismic waves. Currently, seismology concerns itself with the measurements and analysis of all motions which are registered by seismographs on /5
the surface of the solid earth. Day and night, about one thousand seismic stations, scattered in various locations on the globe, record the shaking of the earth's surface arising from various causes. On the earth there is a notable seismic background, or noise, and waves from earthquakes and large explosions--which can be used for scientific purposes in investigating the earth's structure--are recorded along with this noise background. This seismic noise is associated, on the one hand, with industrial activity and transport, and, on the other hand, with microseismics, i.e., with seismic waves which are generated by storms and by the continuous wave-action in the oceans.

In comparison with the earth, the moon is an ideal subject for seismic investigations. This is due to the fact that on the moon there is no atmosphere and no oceans and no industry and, correspondingly, no seismic disturbance. Tied in with this is the fact that the sensitivity of seismometers set up on the moon is outstandingly high and approaches its theoretical limit, amounting to a shift in the ground of some tens of Angstroms. During earthquakes, elastic oscillations--seismic waves--are propagated from an earthquake focus--a limited region under the earth's surface. The region from which the seismic waves propagate is called the earthquake focus. The focus is situated under the surface of the earth, and its extent is equal to several km. Essentially, seismic waves are low-frequency sound waves in the solid, elastic earth. They are divided into body waves and surface waves. The body waves occur as two types: longitudinal and transverse. The longitudinal waves are elastic waves of compression; while the transverse waves are elastic waves of displacement. The propagation of the body waves in the elastic earth is similar to the propagation of light rays in optical media. In particular, the body seismic waves propagate along rays.

The longitudinal and transverse seismic waves, in contradistinction to the surface waves which are propagated along the earth's surface, penetrate the whole volume, the actual body, of our planet. That is why they are called body waves. They, in almost the literal sense of the word, X-ray our planet, for as in X-ray analysis they make it possible to clarify the internal /6
structure of the earth without a direct penetration into its interior. The velocity of the longitudinal waves is about 1.7 times the velocity of the transverse waves. Accordingly, these faster waves are recorded on seismograms earlier and are called primary or P-waves; the transverse waves are named secondary or S-waves. The velocity of the body waves is expressed in terms of the elastic moduli (K--compression modulus, μ--shear modulus) and the density ρ of the medium at a given point, by the following simple formulas, known from any elementary physics course:

$$v_P = \sqrt{\frac{K + \frac{4}{3}\mu}{\rho}} \quad \text{--longitudinal waves} \qquad (1)$$

$$v_S = \sqrt{\frac{\mu}{\rho}} \quad \text{--transverse waves} \qquad (2)$$

Were the velocities of the P- and S- seismic waves in the earth constant and independent of the depth, then the seismic rays would be straight line segments. In actual fact, v_P and v_S systematically increase with the depth of the earth, except for a small zone at depths of 50 to 250 km. Therefore, the true seismic rays are curved; i.e., the earth behaves with respect to seismic rays as a refracting lens (Fig. 1).

According to seismological data, the earth is divided into three fundamental regions: the crust, the mantle, and the core. The crust is separated from the mantle by a sharp seismic boundary, at which the properties of the medium change discontinuously (the velocities v_P and v_S as well as the density ρ increase). This boundary was discovered in 1909 by the Yugoslav seismologist Mohorovičić. Accordingly, the boundary between the crust and the mantle is called the Mohorovičić or M boundary. From this discovery, the earth's crust received a precise definition: specifically, the earth's crust is the external layer of the earth above the M boundary. The thickness of the earth's crust is irregular: it changes from about 10 km (taking into account the depth of the water) in ocean regions to some tens of km in the mountain regions of continental areas. The contribution of the earth's crust to the total mass of the earth and to its moment of inertia is small; therefore, in considering the earth as a whole, it is /7 usual to represent the earth's crust as a homogeneous layer with an effective thickness $\sim$ 35 km. Beneath the crust in the depth interval of 35-2,898 km is located the silicaceous covering or mantle of the earth.[1] Finally, the central part of the earth, located at depths between 2,898 and 6,371 km, forms the core of the earth. Even at the end of the last century, scientists conceived that in the earth there must be a core whose density significantly exceeded the density of the outer silicate mantle. This is how they reasoned about this. The density of the outer minerals composing the crust is equal to approximately 2.8 g/cm³ (granite) and approximately 3.0 g/cm³ (basalt), but the average density of the earth is equal to 5.5 g/cm³, a much higher figure. As a result, the earth must have a heavy core. At the same time, they knew about iron meteorites, with the density of iron under normal conditions being 7.85 g/cm³. These facts served as the basis for advancing the hypothesis of an iron core for the earth.

The seismic boundary at a depth of 2,898 km between the /8 mantle and the earth's core (and thus the earth's core itself) was discovered by the German seismologist, Gutenberg, in 1914. This boundary has no special designation, although one could with complete justification call it the Gutenberg or G boundary. The mantle-core boundary is the sharpest of the boundary divisions in the earth's interior. It strongly reflects the body P- and S-waves and sharply refracts the P-waves. The velocity of the P-waves at this boundary falls discontinuously from a value of 13.6 km/sec in the mantle to a value of 8.1 km/sec in the core; the velocity of the transverse waves correspondingly decreases from

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[1]The terms "covering" and "mantle" are synonyms. In geophysics, the term "covering" is used more often; in geology, "mantle."

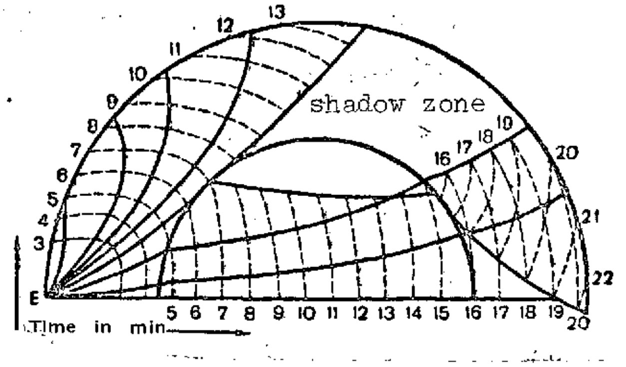

Fig. 1. The paths of the P-waves and their time of propagagion in the earth's interior. This cross-section of the earth shows the paths of the seismic P-waves irradiating from an earthquake focus located directly under the epicenter (point E). The dotted lines--isochrones--show the time of arrival in minutes of the P-waves at various points on the earth's surface. The P-waves are not recorded in the broad shadow zone, caused by the refraction of these waves at the mantle-core boundary.

7.3 km/sec down to zero. Conversely, the density increases from 5.6 to 10 g/cm³. The fact that the earth's core does not allow to pass through itself the transverse S-waves, whose velocity (v_S) in it is equal to zero, indicates that the shear modulus μ of the core is also equal to zero. Consequently, the earth's core appears to be a liquid one. This fundamental conclusion of seismology is also verified by all other geophysical phenomena having to do with the earth's core. Seismological data show that the mantle and core of the earth have a definite "fine" structure. This structure is evident from Fig. 2, in which is shown the seismic model of the earth, that is, the distribution of velocities (for both P- and S-waves) as a function of depth.

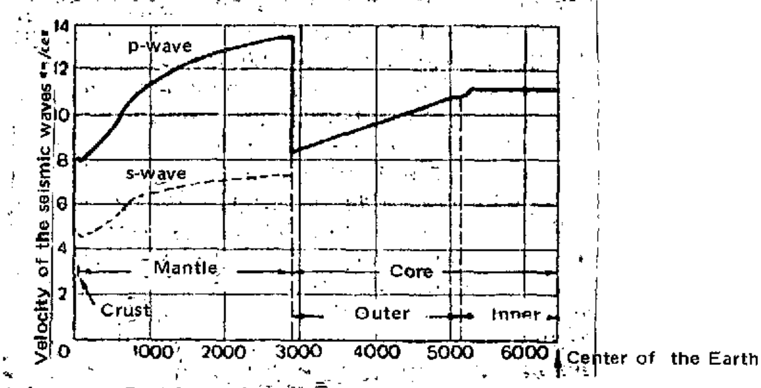

Fig. 2. Velocities of the P- and S-waves inside the earth. The seismic model of the earth.

In accord with seismological data, the earth's interior is divided into eight characteristic regions or zones. These zones are designated by upper-case letters of the Latin alphabet: A, B, C, D (D', D"), E, F, G. Zone A (0 to 33 km) is the earth's crust; zone B (50 to 350 km) is the subcrustal zone, a layer of lowered velocities; zone C (350 to 1,000 km) is a transition layer, a zone of anomalously rapidly increasing velocities for P- and S-waves; zone D is divided into zone D' (1,000 to 2,700 km)--a zone of normally increasing velocity as a result of the pressure of the overlying layers--and zone D"--a narrow boundary zone of the mantle with the core, characterized by a constant velocity for the P- and S-waves; zone E (2,900 to 4,980 km) is the liquid outer core; zone F (4,980 to 5,120 km) is the transitional zone of the core with a complex velocity profile; and zone G (5,120 to 6,471 km) is the solid inner core of the earth. Only very recently, the precision of seismological observations has been remarkably raised as a result of going from observations at isolated seismic stations to observations at "seismic profiles" containing hundreds of seismic stations arrayed along definite directions. As a result, indications have appeared of the existence of breaks in the velocity distributions of the P- and S-waves in the earth's mantle. With the greatest precision, two boundaries have been shown: one in the depth interval of 300 to 400 km, and the other at the depth interval of 600-700 km. Such large indeterminacy is related to the fact that the seismology of these depths is extraordinarily complex and intricate. Finally, the latest achievement of seismology is the observation of weak reflected waves from the boundary of the inner core, which seems to be the first direct evidence that the inner core is in the solid state, in contrast to the outer core, which is in the liquid state. /9

The properties of the velocity distribution of seismic waves in the earth's interior is explained as follows. In going from the earth's crust (granite, basalt) to the mantle (ultrabasic rock), the velocities increase abruptly. In the subcrustal zone there is a layer of lowered velocities which is associated with the proximity of the inner temperature in this layer to the melting points. In zone C, the velocity quickly rises as a result of the phase transitions of the minerals into denser, mechanically harder modifications. Then follows the homogeneous layer D, where the velocity grows only as a result of squeezing from the pressure of overlying layers. At the boundary with the core, there is a small velocity plateau, whose appearance even now has not been completely explained. The fall in the velocity of the P-waves in going from the mantle to the core is related to the fact that the core is liquid and consists of denser material. It has turned out that the compression moduli K for the mantle and the core at their boundary are approximately equal, while the density of the mantle ρ_M (2,898) = 5.6 g/cm³, significantly less than the density of the core ρ_C (2,898) = 10 g/cm³. /10

The earth's core consists essentially of iron and a small admixture of light elements. Most probably, the substance of the earth's core contains as an impurity silicon and perhaps sulfur, but this question still remains debatable. The increase in the velocity of the P-waves in the outer core has a normal appearance and is brought above by the pressure increase towards the earth's center. Determining the detailed change of the velocity in the transition zone F of the core still remains an unsolved problem. What is clear is only the fact that the velocity in this zone of the earth rises and that this rising itself is brought about by a transition of the material from the molten state to a solid, crystalline state. The velocity of the longitudinal waves in the inner core practically does not change since the pressure in this region of the earth rises very weakly.

Up until now, we have concentrated all our attention on the body seismic waves and have only mentioned the surface waves. These surface waves are widely used to investigate the outer layers of the earth (crust and outer mantle). Surface waves, like body waves, occur as two types and have received the names Rayleigh waves and Love waves, after those scientists who discovered them theoretically. These waves were studied theoretically by Rayleigh in 1885 and by Love in 1911. All seismic waves on the seismogram were discovered at the very end of the last century, when the recording of the surface seismic Love waves was not understood, until the publication of the theoretical work of Love in 1911. In the generation of a Rayleigh wave, the displacement of the soil particles is in the vertical plane, and they describe counterclockwise ellipses. This motion of the sections or segments in the wave occurs almost as if they were rolling towards the wave source. In contrast to these Rayleigh waves, in /11
the Love waves the shift of sections is in the horizontal plane, perpendicularly to the direction of wave propagation. In the surface waves, the amount of displacement is maximal at the surface and rapidly (exponentially) decreases with increasing depth. Apropos of this, with the aid of surface waves, it is possible to effectively study (sound or probe) the earth to depths equal to about a third of their wavelengths. The length of surface waves initiated by earthquakes lies in the interval from tens to many hundreds of km. Therefore, it is possible to investigate the outer layers of the earth to a depth of hundreds of km with the aid of surface waves. The surface waves from especially strong earthquakes are so intense that they travel around the globe several times. Such intense waves make it possible to get much information about the interior of the planet without the use of a large number of instruments. Consequently, they are very convenient during seismic probings of the moon and planets. Many interesting results have been obtained with the aid of surface waves. They have made possible a rather detailed study of the distribution of the layer of lowered velocities in

the outer mantle of the earth and of the structure of the earth's crust of the continental and of the oceanic type, and a series of other regional details of the outer layers of the earth. However, in this brochure, we are interested in the question of the global structure of the earth and planets. Therefore, we shall not get into detail concerning these preceding results. Also, the limitations of space do not permit us to speak about the earth's natural oscillations.

Gravimetry

Gravimetry is second in significance in the realm of geophysics, yielding only to seismology. The gravitational field of the earth reflects the nature of the mass distribution in the interior of our planet and is closely related to the shape of the earth. The practical value of gravimetry is enormous. On the one hand, gravimetry is related to the shape of the earth, and, thus, to geodesy (large-scale surveying) and this, in turn, is related to topography. On the other hand, the gravitational field determines the external ballistics of the earth, whose significance over eons needs no comment. In this broad context, /12
gravimetry (the science of the gravitational field and the shape of the earth) is the oldest geophysical discipline.

That the earth is spherical was guessed even in the depths of antiquity and the first determination of the earth's radius was completed by a scientist from Alexandria, Eratosthenes, about 235 B.C. However, naturally, gravimetry as a science could not develop until the law of universal attraction was discovered. Newton in the third part of his "Mathematical Principles of Natural Philosophy" set forth a theory of the shape of the earth based on the law of universal attraction. Newton first understood that, as a result of the rotation of the earth, its shape must be not a sphere but rather an ellipsoid of revolution. Consequently, the earth is flattened at the poles and elongated in the equatorial region. Newton first calculated the oblateness of the earth:

$$\alpha = \frac{a - b}{a} \tag{3}$$

where a is the equatorial radius;

b is the polar radius of the planet.

It is true that the number which he obtained, $\alpha = 1/230$, is not quite correct. The modern value for the oblateness of the earth has been determined with great precision and is equal to $\alpha = 1/298.25$.

It is interesting to note that the conclusion of Newton about the oblateness of the earth was contradicted by many scientists, among whom was the noted French astronomer G. D. Cassini. In connection with this problem, the French Academy of Sciences in the middle of the eighteenth century organized expeditions to carry out angular measurements at various latitudes. As a result of the measurements made, it was shown that the shape of the earth is a flattened spheroid with a polar axis about 20 km shorter than the equatorial axis. The point of view of Newton about the spheroidicity of the shape of the earth received experimental verification and thus won out.

Modern gravimetry takes its origin from the remarkable work of the French mathematician Clairaut, "The Theory of the Shape of the Earth, Based on the Principles of Hydrostatics," published in 1743. Relying on the law of universal attraction, Clairaut rigorously showed that the gravitational acceleration on the /13
surface of the earth's spheroid as a function of latitude changes in accord with the following simple law:

$$g = g_{\rho}(1 + \beta \sin^2 \varphi) \qquad (4)$$

where φ is the latitude of the place; g_{ρ} is the gravitational acceleration at the equator; $\beta = (5/2)q^{\rho} - \alpha$; α is the oblateness; $q = \frac{\omega^2 a}{g_e}$, the ratio of the centrifugal force to the force of gravity at the equator; ω is the angular velocity of rotation of the earth; and α is its larger semi-axis.

We see that the theory of Clairaut, incorporated into equation (4), leads to a completely new status in the question of the shape of the earth. It makes it possible to determine the oblateness α independently of measuring geometric elements by way of angular measurements. According to the theory of Clairaut, in order to determine the oblateness α of a planet, it is sufficient to determine the gravitational field on its surface. Consequently, gravimetry as a geophysical discipline studies the force of gravity and its distribution over the surface of the earth and determines the shape of the earth from this measured distribution.

Further development of gravimetry (or, as they now more often call it, the theory of the shape of the earth) was associated with the work of the English physicist D. G. Stokes and the Soviet geophysicist and correspondent member of the Soviet Academy of Sciences M. S. Molodenskiy. The significance of gravimetry for the study of the internal structure of planets is tremendous. However, for the planets seismic data is lacking. Nevertheless, many planets possess natural satellites. Observation of these natural satellites permits the gathering of data about the gravitational field of the planet and, thus, data about the mass distribution in the interior of the planet and the planet's oblateness. Data on the gravitational field of planets coupled with the values of their average densities is the only observational information about planets which is used in building models of their internal structure. In this chapter, we are forced to introduce some fundamental formulas. These formulas are essentially elementary, and no matter how brilliant the exposition, without the use of these fundamental relationships, even in the best case it would only create the illusion of understanding the subject matter.

The Gravitational Field and the Shape of the Earth. The Earth's Moment of Inertia

If the earth were an exact sphere, in which the density distribution depended only on the radius, i.e. $\rho = \rho(r)$, that is, if it were spherically symmetrical, then the external gravitational potential of the earth would have the following extraordinarily simple form: /14

$$V = -\frac{G \cdot M}{r} \tag{5}$$

where r is the distance from the center of the sphere; G is the gravitational constant; M is the mass of the planet. The gravitational potential or the gravitational potential energy both mean the same thing and they mathematically describe the gravitational field. The real earth is approximately a sphere. It deviates from a sphere by one three-hundredths or less. Even though this deviation is small, it deserves consideration since it contains valuable information on the small fluctuation of density in the earth's interior, the differences in moments of inertia of the earth relative to its major axis, and the deviation of the earth's interior from the state of hydrostatic equilibrium. Even before the launching of artificial earth satellies, scientists succeeded in determining the first correction term J_2 to the Newtonian part of the gravitational field as given by (5). As a result, the external gravitational field of the earth is represented by the formula:

$$V = -\frac{GM}{r}\left[1 - \left(\frac{a}{r}\right)^2 J_2 P_{20}\,(\cos\Theta)\right], \tag{6}$$

where α is the equatorial semi-axis;

$$J_2 = \frac{C-A}{Ma^2} \quad \text{--the gravitational moment;} \qquad (7)$$

$$P_{20} = \frac{3}{2}\cos^2\Theta - \frac{1}{2} \quad \text{--the second Legendre polynomial} \qquad (8)$$

Here C is the moment of inertia relative to the polar axis; A is the moment of inertia relative to the equatorial axis; θ is the polar angle equal to the complement of the latitude, that is, $\theta = \frac{\pi}{2} - \phi$. The modern value for J_2 has $J_2 = 1{,}082{,}65 \cdot 10^{-6}$. Thus, the value for J_2, which characterizes the deviation of the gravi- /15 tational field of the real earth from the spherically symmetric part (5), has turned out to be, as it should, on the order of magnitude of the oblateness of the earth, equal to one three-hundredth. In this regard, the oblateness α of the terrestrial spheroid is related in a simple fashion to J_2, to the angular velocity ω of the earth's rotation, to the total mass M, and to the equatorial radius α:

$$\alpha = \frac{3}{2} J_2 + \frac{1}{2}\frac{\omega^2 a^3}{G \cdot M}$$

If the whole surface of the earth were covered by a global ocean and its surface were not disturbed by wind-driven waves and tides, then the shape of the earth would coincide with the shape of the terrestrial spheriod.

For the problem of the internal structure of the earth, of first-rate interest is the value of the average moment of inertia

$$I = \frac{C + 2A}{3} \qquad (9)$$

which, coupled with the value for the average density

$$\varrho_0 = \frac{3M}{4\pi a^2 b}$$

and some seismological data, makes possible the determination of the density distribution in the earth's interior.

In order to determine I, it is necessary to know along with J_2 some other quantity somehow or other related to the moments of inertia C and A. No success has been achieved in trying to get from purely gravimetric measurements still one more relationship between the moments of inertia C and A. However, here astronomy comes to the aid of gravimetry, for its methods make possible the determination of the precession constant of the earth's axis:

$$H = \frac{C - A}{C} = 0.0032732$$

The density distribution in the interior of a planet greatly influences the average moment of inertia (9) and, conversely, the value of I--determined experimentally--essentially controls the density distribution in various model calculations. Let us consider the case of a homogeneous model--a planet with a constant density distribution. Evaluating the moment of inertia of such a homogeneous sphere presents no difficulty. As a result we have /16

$$I^* = \frac{I}{MR^2} = 0.4 \qquad (10)$$

Thus, we come to the simple but important conclusion that, in the case of a planet with a constant density, its dimensionless moment of inertia I* is equal to four tenths. It is easy to convince oneself by means of direct numerical calculations that with an increase in density in a planet's interior, from the periphery to the center, the quantity I* will take on values less than four tenths. On the other hand, if with increasing depth the planet diminishes in density, i.e. a lessening of density with depth takes place in the planet, then the value of I* will exceed the limiting value of four tenths. For the earth, the value for I* is equal to 0.3309 according to observations. This figure corresponds to a quite substantial concentration of mass in the central regions of the planet. Appreciable gravitational fields are at work in the interior of planets. Therefore, if for some reason or other in the evolution of a planet an inversion of densities occurs in its interior, that is, zones of lowered density are situated under regions of greater density, then there arise powerful Archimedian forces striving to swap the locations of these regions. In such a situation, it is said that the condition of mechanical equilibrium in the planet has been destroyed. Therefore, the density is an increasing function of depth and its increase arises as the result of compression under the pressure of the overlying layers, as a result of the increase with depth of the concentrations of heavy components, and sometimes as the result of a density increase attendant upon phase transitions under high pressures.

In the deep interior there also exist processes leading to a lowering of density. The fundamental ones of these are: heating (the increase in temperature), melting, partial or fractional melting with the separating out of components with lower density (for example, the melting out of basaltic magmas in the interior of the earth and in the interior of the moon). As a rule, however, the processes leading to a lowering of density are less effective than the causes forcing the density to grow with depth. An external manifestation of the fact that on a global scale the density increases with depth or in the case of small bodies remains almost constant is the condition that $I^* \leq 0.4$.

Investigation of the gravitational field of the moon with the aid of artificial moon satellites has permitted determination of its dimensionless moment of inertia /17

$$I^* = 0.402 \pm 0.02$$

This fundamental result shows that the density of the moon is approximately constant. From the physical point of view, this deduction is natural: the pressure at the center of the moon does not exceed 50,000 atm and, accordingly, the increase in density as a result of pressure amounts to only some few percent.

It is interesting to recall the history of the determination of I^* for the moon prior to the launching of artificial satellites. About ten years ago, a well-known American astronomer--Eckhardt--undertook an attempt to determine I^* for the moon by way of a detailed analysis of librational oscillations of the moon in its orbital motion around the earth. He obtained a value for I^* notably above the limiting value 0.4. The work of Eckhardt served as a basis for the assumption of an anomalous density distribution in the moon's interior, specifically for the assumption that there is a remarkable fall in the density as a function of depth. Such a strange but effective result contradicted common sense and forced people to think that the result of Eckhardt was erroneous. As we now know, these misgivings turned out to be correct, and, currently, the value of I^* for the moon does not bring about any quandaries.

The External Gravitational Field of the Earth According to Data from Artificial Earth Satellites

Prior to the launching of artificial satellites, the external gravitational field of the earth was described by the simple two-term formula (6). It would be incorrect to think that the gravitational field of our planet would be so simple. In actuality, the simplicity of the earth's gravitational field was related to the fact that no one succeeded in covering the earth with a detailed network of gravimetric surveys which should have made possible the production of further corrections to the basic Newtonian part of the field (5).

In the general case, the field of any attracting cosmic body--planet, satellite, or star--can be decomposed into a series of spherical functions. These spherical functions always enter the picture when any problem is being solved for a sphere or for a body whose shape is approximately spherical. They have a definite form as a grouped sum of cosines and sines of angular variables, i.e., of the polar distance (or latitude) and the longitude. These spherical functions are the so-called eigen- functions for a sphere and hence their very great value to geo- physics. In solving any problem, the choice of functions in which this problem is solved is dictated by conceptual conveni- ence. The eigenfunctions for a given problem always are the most natural and convenient and simple. /18

Since the earth is quite close to being a sphere--it is almost a sphere--one has to deal with spherical functions in geophysics for practically all the problems. As we have said, the gravitational field of the earth is decomposed into spherical functions; the magnetic field of the earth has been decomposed into spherical functions ever since the time of the great German mathematicicn Karl Gauss; the free and the natural oscillations of the earth are also decomposed into spherical functions. This decomposition into spherical functions is called "spherical analysis." Currently, scientists have subjected to spherical analysis the profile of the earth's surface and the moon's sur- face; the heat flow from the earth's interior; and other geophy- sical fields.

Unheralded, we have already met with our first spherical functions in this brochure. As we know, expression (6) gives the first terms in the decomposition of the gravitational potential. Consequently, this is the start of the series for the decomposi- tion of the potential into spherical functions. Actually the simplest spherical function is 1--the spherical function of zero degree. The spherical function of first degree consists of three components: $\cos\theta$, $\sin\theta\cos\lambda$, $\sin\theta\sin\lambda$ (θ is the polar angle and λ is the longitude; these then are the angular coordinates in a spherical system of coordinates). The decomposition of the gravitational potential does not contain components of the spher- ical function of the first degree. This is related to the fact that we have made a judicious choice for the origin of our coor- dinate system; specifically, we have placed it at the earth's center of mass. The spherical function of second degree consists of five components. One of these components, specifically P_{20} (6), (8), enters into the second member of the decomposition of the potential (6). Once more, we eliminated in (6) the other components (of the second-degree spherical function) and thereby got a simpler and more convenient expression for the potential as the result of a judicious choice of the coordinate axes: the /19
coordinate axes coincide with the major axes of inertia for the

planet. In the general case, the spherical function of the nth degeee contains (2n + 1) components and the decomposition of the earth's gravitational potential has the following form:

$$V=\frac{GM}{r}\Big\{1-\sum_{n=2}^{\infty}\Big(\frac{a}{r}\Big)^n J_n P_n(t)+ \\ +\sum_{n=2}^{\infty}\sum_{m=1}^{n}\Big(\frac{a}{r}\Big)^n P_{nm}(t)\,(A_{nm}\cos m\lambda+B_{nm}\sin m\lambda)\Big\}. \tag{11}$$

Here r, θ, and λ are the spherical coordinates at the point of observation; $t = \cos\theta$; P_n is the Legendre polynomial of the nth degree and is a polynomial of the nth degree in cos θ. P_{nm} is the associated Legendre polynomial, a polynomial of the nth degree in cos θ and sin θ; J_n, A_{nm}, and B_{nm} are gravitational moments determinable experimentally from trajectory observations of artificial satellites. The components of the spherical functions of the nth degree entering into (11) have the following forms:

$$P_n(\cos\theta);\ P_{nm}(\cos\theta)\cos m\lambda;\ P_{nm}(\cos\theta)\sin m\lambda;$$

$$m = 1, 2, 3, \ldots, n-1, n$$

The remaining symbols in (11) are the same as in (5) and (6) and are standard.

Prior to the launching of artificial satellites, only one coefficient J_2 in the decomposition (11) had been determined, and this required completion of a grandiose scheme to cover the earth with geodesic and gravimetric surveys. Quite recently, the question of determining the other coefficients in the decomposition of the earth's potential was viewed quite skeptically. Thus, the outstanding geophysicist of the first half of the twentieth century, Harold Jeffries, wrote in 1959 in his classic monograph "The Earth" that it was possible that the coefficient J_4 would be determined within 20 years provided there was no slackening of the pace of astronomical and geodesic work. Jeffries thought that the coefficient J_3 would be much less than J_4 and that therefore the next correction term for the two-term potential (6) would be the expression with J_4. We shall say more about this mistake of Jeffries later.

The wide use of artificial satellites for geodesic purposes has radically changed the situation. Observations of satellites (including special geodesic ones) with the aid of modern optical and radio-astronomical instruments plus the use of computers for /20 data processing have already made possible by the beginning of the '60's the determination of about 10 zonal moments J_n and some tens of tesseral moments A_{nm} and B_{nm}. The zonal moments J_n in the decomposition of the potential (11) bring about the secular perturbations of the orbits of artificial earth satellites. Therefore, in the determination of J_n, one uses a comparatively lengthy series of observations; these moments are determined with more precision than are the tesseral moments A_{nm} and B_{nm}, which bring about only short-term changes in the elements of the orbits.

The determination of the gravitational moments with the aid of artificial satellites belongs to the most brilliant pages of the history of geophysics and even of science. One can be so bold as to place this result beside such achievements as the discovery of the radiation belts and magnetosphere of the earth. We shall tell below what important conclusions follow from the detailed investigations of the earth's gravitational field with the aid of artificial satellites.

Deviation of the Earth from the Condition of Hydrostatic Equilibrium

A geophysicist as remarkable as Jeffries, who figured erroneously that the gravitational moment J_3 is much less than J_4, had weighty reasons for such a supposition. Presumably, he reasoned approximately along these lines. Everything indicates that the earth is in a state close to hydrostatic equilibrium. If you study the decomposition of the earth's field into spherical functions (11), it is possible to obtain some quantitative features of the deviation of the earth from the state of hydrostatic equilibrium. At first let us assume that the earth is exactly in a state of hydrostatic equilibrium. We pose the question: what form of the gravitational potential (11) will correspond to this assumption that we have made? The answer to this is easy. With hydrostatic equilibrium, the expression for the potential has the following form:

$$V = \frac{GM}{r}\left\{1 - \left(\frac{a}{r}\right)^2 J_2P_2(t) - \left(\frac{a}{r}\right)^4 J_4P_4(t) - \left(\frac{a}{r}\right)^6 J_6P_6(t) - \dots\right\},$$

that is, it contains only the even zonal moments J_{2n}, while the odd zonal moments J_{2n+1} and all the tesseral moments A_{nm} and B_{nm} are equal to zero. But this is not all. Under hydrostatic equilibrium, the values of the even zonal moments J_{2n} must decrease very rapidly with increasing *n* in accord with the following law: /21

$$J_2 \sim \frac{1}{300},\ J_4 \sim \left(\frac{1}{300}\right)^2,\ J_6 \sim \left(\frac{1}{300}\right)^3,\ \ldots,\ J_{2n} \sim \left(\frac{1}{300}\right)^n.$$

Geophysicists knew that the earth is in a state approximating hydrostatic equilibrium. In this case, it was completely natural to assume that the correction term to the first two components in the formula for the gravitational potential (6) would involve J_4. And this is how the majority reasoned, until the advent of the satellite measurements. Well, what did the satellite data show? Essentially, the data gave a sensational result, namely: all gravitational moments beginning with J_3 are approximately of the same order of magnitude, equal to some units multiplied by 10^{-6}, that is, all moments except J_2 turned out to be quantities on the order of the square of the oblateness of the earth and, furthermore, the decrease in the moments with increasing n takes place significantly more slowly than was expected. Thus, as a general fundamental conclusion from the satellite data, we state that the deviation of the earth from hydrostatic equilibrium is on the order of the square of its oblateness. Let us explain this conclusion in a physically more concrete way. The deviation of the state of the earth from hydrostatic equilibrium means that in the earth tangential strains are operating along with hydrostatic strain or pressure. It is possible to evaluate the tangential strains to within an order of magnitude in the following way. The deviation of the earth from equilibrium by an amount on the order of magnitude of the oblateness shows that the shape too of the earth deviates from equilibrium shape by the same order of smallness. In order to get a measure of the thickness of the nonequilibrium stratum, we must multiply the square of the oblateness α^2 by the average radius R of the earth. As a result we get a stratum with a thickness of 70 m. One can then calculate that the tangential strains in the earth's interior that arise from such a stratum are equal to some tens of kg/cm^2. The detailed distribution of strains in the earth's interior has not been /22
successfully established, since this is a very complicated undertaking.

Isostasy

Knowing the detailed structure of the earth's gravitational field, established with the aid of satellite data, one can establish just as detailed a picture of the deviations of the shape of the earth from equilibrium shape. We already know that these deviations are on the order of the square of the oblateness of the global spheroid and, in linear units, are on the order of tens of meters.

Here it will be pertinent to explain in more detail the terminology accepted in geophysics, in that branch of it which treats the shape and gravitational field of the earth, which have only been superficially touched upon in this chapter. The topographic surface of the earth is extremely irregular. Therefore, in geophysics, under the heading of "the earth's shape" is placed a shape with a certain conventional surface approximating the surface of the real earth. If the earth were a liquid rotating planet, then to determine its shape it would be sufficient to know the expression for the external gravitational potential ("geopotential") W, which is made up of the gravitational potential V as given by (11) and the centrifugal potential arising from the rotation of the earth. Then the surface of the earth would be a level surface and its equation would be determined in the usual way: $W = K_0$, where K_0 is the value of the external potential at the surface of the planet. In this determination, the shape of the planet is related to such physical parameters as the mass distribution within the planet and the planet's angular rotational velocity. Therefore, even though the earth is not in hydrostatic equilibrium, in geophysics the shape of the earth is determined with the aid of the condition $W = K_0$; this shape is called the geoid. Three fourths of the earth's surface is covered by oceans. Naturally, the surface of the oceans unperturbed by wind currents coincides exactly with the surface of the geoid, while on dry land the geoid is located under the surface of the continents. As we have said at length above, the gravitational field and, correspondingly, the geopotential are made up of components notably different in their value. In regard to this, the geopotential W is divided into two parts: a major part and a correction. The major term contains the Newtonian potential, the first correction term (that is proportional to J_2 (6)), and the /23
centrifugal potential. This combination is called the normal field W_0. The correction part of the geopotential contains all the remaining terms (which are on the order of the square of the oblateness) and is called the perturbation T.

By analogy with the division of the external field W into a normal field W_0 and a perturbation T, the geoid is built up in two ways. First the basic reference shape is determined, i.e., the normal shape; then the heights (small in value) on the geoid are determined, i.e., the distances or deviations of the geoid from the normal shape. At first glance, it is possible to get a good approximation if we take for the normal shape the Newtonian sphere with an average radius R and average density $\bar{\rho}$. Since the deviation of the potential W from the Newtonian potential (GM/r) (5) is on the order of the oblateness $\alpha = 1/300$, then the average height of the geoid above the sphere will be on the order of $\alpha \cdot r \approx \frac{6.4 \cdot 10^3}{300} \approx 21$ km. This value is small in comparison with the dimensions of the earth, but it is large compared with the typical heights of the earth's relief. Therefore, one chooses for the normal shape an ellipsoid of rotation which is an equipotential surface for the normal potential W_0. It is just this shape that we have been generally calling heretofore the earth's spheroid. This ellipsoid is sometimes called the reference ellipsoid.

The normal ellipsoid is quite a good approximation to the geoid. Actually, the external potential deviates from the normal one by a value on the order of α^2. Consequently, the deviation of the geoid from the normal ellipsoid (the heights of the geoid) is on the order of $\alpha^2 R \approx 70$ m. With the aid of satellite data, it is easy to construct maps of the heights of the geoid. Such maps have been constructed. The heights of the geoid quantitatively characterize the deviation of the earth's gravitational field from the normal field. In principle, it would be possible to assume that the deviation of the true gravitational field from the normal one is due to the relief of the earth. Thus, in places where there are mountains, the gravitational field is stronger because of the supplementary attraction by the mountains, and in places where there are depressions the field is weaker because of the deficit in mass. In actuality, the map of the heights of the geoid show that these deviations are not related to the major topographical features (oceans and continents) of the earth.

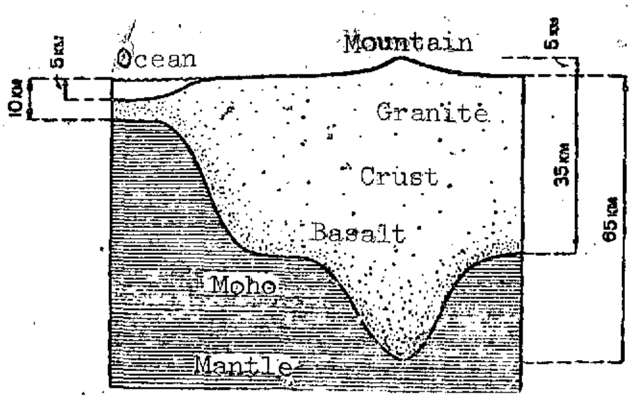

Fig. 3. The isostatic equilibration between the crust and the mantle.

From this we draw the most important conclusion that the continental regions are isostatically compensated: the continents /24 are floating in the subcrustal substrate like giant icebergs in the polar seas. Just slight deviations of the earth's gravitational field from the normal are brought about by some fluctuations of the density in the crust and in the mantle of the earth.

Qualitatively, the idea of isostasy was introduced into geophysics in the middle of the last century. This concept was advanced to explain the surprising fact that the presence of mountains exerts almost no effect on gravimetric measurements. According to the principle of isostasy, the light crust consisting of granite and basalt is isostatically equalized or equilibrated on the heavier mantle, as is shown in Fig. 3. It is accpeted that the mass of a substance per unit area measure down to some standard level surface in the interior is approximately the same for all surfaces of the earth. We see that the light substance of the earth's crust, if it happens to form somewhere a mountain system, is submerged to larger depths in the heavy rocks of the mantle. To describe such a situation, it is figuratively said /25 that the mountains have roots penetrating into the depths.

The presence of isostasy leads to important properties in the structure of the earth's outer layers. These properties, shown in Fig. 3, are verified with the aid of detailed seismic investigations.

The investigation of the earth's gravitational field with the aid of artificial satellites has made it possible to characterize quantitatively with significantly greater detail the isostatic compensation of the crust for all planets.

As we have already said, it is accepted that the earth's crust is floating, as it were, on the underlying or basement rock of the mantle. However, according to seismological data the transverse seismic waves (S-waves) pass through the mantle and, thus, the mantle must be in the solid state. What is going on here? Here is the answer. For periodic oscillations with periods on the order of seconds, hours, or days (corresponding to body and surface seismic waves, the natural oscillations of the earth, tides), the mantle behaves like an elastic solid. For motions, though, with periods on the order of ten thousand years, the material of the outer mantle flows like a liquid. The liquid, with a relaxation time on the order of ten thousand years and with the mechanical parameters of the outer mantle, must have a very high viscosity, on the order of 10^{21} poise. A substance having such properties will flow under loads acting for thousands of years and will react like an elastic solid to periodic processes in the range from seismic waves to tides.

Geomagnetism. Distribution of Electroconductivity

Geomagnetism is one of the oldest and broadest of the geophysical disciplines. Yet for many years, in courses on the internal structure of the earth, the problems of geomagnetism were not handled. Such a situation, at first glance paradoxical, had a quite simple, even trivial explanation. Geomagnetism added nothing to what was known about the interior of the planet; and the theory of the earth's magnetic field had a rather formalistic character. It said nothing about the physical causes for the /26 origin and maintenance of the earth's magnetic field during cosmic intervals of time. Magnetic fields are widely scattered throughout the universe. They exist in stars, in outer space, and even the sun and the planet Jupiter have a magnetic field. It is not beyond the realm of possibility that a magnetic field will be discovered eventually for Saturn, Uranus, and Neptune.

In the problem of the internal structure, perhaps the most remarkable property of the geomagnetic field is its rapid variability. The significance of the variations of the magnetic field for the physics of the earth comes from the fact, on the one hand, that they are the most rapid changes of all the geophysical processes which are subject to study and from the fact that, on the other hand, they reflect the complex picture of the hydromagnetic currents and oscillations in the earth's core--the place where the sources of the earth's own magnetic field are located. This also makes it possible for us to make judgments about the values of a series of parameters for the earth's core, which values cannot be evaluated with the aid of other geophysical methods. Furthermore, the study of the attenuation in the earth of electromagnetic signals which are initiated by solar activity in the outer atmosphere permits the determination, at least in coarse detail, of such important characteristics of the earth's interior as its electroconductivity.

Along with the two applications to the physics of the earth already mentioned, geomagnetic investigations now are widely used for establishing what movements have occurred in the earth's crust and oceans during historical and geological times. The latter has become possible on the basis of recently worked out archeomagnetic and paleomagnetic methods which permit the determination of the geomagnetic field in the distant past. For example, the lava from a volcano takes on a magnetization at the time of its cooling in the earth's magnetic field, and such a magnetization is directed parallel to the applied field. The age of a rock can be determined with the aid of radioactive or geological dating. Consequently, this "fossilized" magnetism lets us make judgments about the magnetic field of the earth which existed at the time of the solidification of the lava. Experiments carried out show that the earth had a magnetic field at least hundreds of millions of years ago and maybe more and that during the course of geological history a change in the polarity or direction of the field occurred. /27

The fact that the earth has a magnetic field was known even in antiquity, more than a thousand years ago, to the Chinese, who were acquainted with the magnetic needle-compass. However, the beginning of geomagnetism as a scientific discipline is set significantly later, in 1600, when William Gilbert, court physician to Queen Elizabeth I, published his treatise on geomagnetism. Gilbert showed that the magnetic field of the earth is similar to the field of a magnetic dipole; that is, the earth is a gigantic magnetic arrow in the shape of a sphere. Magnetologists later discovered that the earth's field is similar to the field of a spherical magnet whose axis is inclined to the axis of the earth's rotation by 11°.

Systematic observations on the magnetic field were begun in 1580 and this permitted in 1622 the discovery of a noticeable change in the direction of the magnetic field in the region of London after the lapse of 40 years. The fundamental work of Gauss, "The General Theory of the Earth's Magnetism," appeared in 1839. Gauss was the first to carry out a spherical analysis of the geomagnetic field, that is, he decomposed the magnetic field of the earth into spherical functions. We have already spoken in detail in the chapter devoted to gravimetry about the fact that, in geophysics, in dealing with the sphericity of the earth, all fields are decomposed into spherical functions. Such an analysis always produces very much and, accordingly, Gauss obtained right away many fundamental results. First of all, he completely unambiguously divided the total geomagnetic field into an internal one and an external one; that is, into fields whose sources were located within and outside of the surface of the earth. As we now know, the sources of the internal field are in the earth's core, except for a small background from the magnetization of the rocks. It has turned out that almost all the earth's magnetism has sources from within the earth. The spherical analysis essentially gives a decomposition of the field into components of various multipolarity: a dipole field; a quadrupole field; and components of higher multipolarity. Gauss showed that, in the natural magnetic field, the dipolar component dominates, but that this component does not account for the entire magnetic field of the earth. The earth also has a quadrupole magnetic field and a field of higher multipolarity. Having decomposed the field into spherical functions, Gauss first calculated the value of the magnetic dipole, which for that period of time was equal to $8.5 \cdot 10^{25}$ cgs units. It turned out that the dipole field of the earth exceeded by an order of magnitude the magnetic field from the higher multipoles. /28

After Gauss, the spherical analysis of the magnetic field of the earth was carried out repeatedly. Elements of the magnetic field began to be depicted with the help of isolines on magnetic charts for various epochs. A detailed comparative analysis of the magnetic charts for various epochs was carried out in 1950. This analysis led to an important discovery. It was accepted that the pictures of the isolines for the non-dipolar part of the magnetic field showed a systematic shift--a drift--in the westerly direction. The amount of this drift is great and amounts to about 2° of longitude per year. Since the sources of the field are located in the liquid core of the earth, this phenomenon then means that in the core there are taking place longitudinal flows of liquid with a velocity of about 0.1 cm/sec. These velocities are a million times greater than the velocities of the tectonic motions which lead to mountain formation and about which we can also make judgments by current shifts in the earth's surface. These latter velocities amount to only about 0.1 to 1 cm per year and less. To get a more graphic representation of the scale of tectonic velocities, let us assume that a particle or segment of the earth's mantle is moving upwards along a radius from the boundary with the core towards the earth's surface with a velocity of 1 cm/year. Then the whole trip of 2,900 km will take 290 million years. If the velocity of the movement is less, say 0.1 cm/year, then the time to exit at the surface will be about 3 billion years; that is, close to the amount of time for the existence of the earth as a planet. The movements in the earth's core are rather faster and it is evident that the field with its secular variations moving westward completes a round trip around the earth's axis in about 2,000 years.

The value of the earth's magnetic field also does not remain constant. Since the time of Gauss, it has been systematically decreasing. At such a rate of decrease, the earth's dipolar magnetic field should disappear within 2,000 years. However, such extrapolations in geophysics are dangerous. No one can say with certainty whether or not epochs of diminishing magnetic field alternate with epochs of its increase. Such an alternation does take place in the vertical movements of the earth's surface. We know that often periods of uplift of the surface alternate /29
with periods of sinking, such that the motion has a rather oscillatory character. Probably, the change in the magnetic dipole of the earth also has an oscillatory character. One should not yet draw final conclusions on this question, since the interval of time during which observations of the magnetic field have been conducted is still too short.

The question of the causes of magnetism in cosmic bodies and, in particular, in the earth has attracted the attention of many scientists. In the last decades, these scientists have probed the correct ways that would, in principle, make possible the explanation of geomagnetism, but a definitive theory still has not been propounded. Propounding a theory of geomagnetism remains one of the unsolved fundamental problems of geophysics.

Modern theories of geomagnetism arise from the assumption that the earth's magnetic field is created and maintained by the so-called dynamo mechanism. Roughly, it is assumed that the creation of the magnetic field in the core arises just as it does in the dynamo in a state of self-excitation. The operating principle of the dynamo is as follows. Let a conducting coil rotate in an external magnetic field. Then, as a result of electromagnetic induction, an electric current arises in the coil. The electric current creates a magnetic field, which can strengthen or reinforce the external magnetic field which, in turn, strengthens the current in the coil, and so forth.

The liquid core of the earth is not at all similar to a real dynamo. But, in principle, if any thermal or gravitational convection currents arise in the liquid conducting core for any reasons, then some system of hydrodynamic flows is established. Thus, we have some system of flows of a conducting fluid. The flow of a conducting fluid in this analogy that we are considering corresponds to the motion of a conductor. If there are any priming magnetic fields in the core, then, when the lines of force of these fields are crossed by a conducting stream, an electric current arises in this stream. The electric current creates a magnetic field, which with appropriate geometry of the flows can strengthen the external priming field, and this, in turn, strengthens the current and so forth. This process will continue until a stationary magnetic field arises and the various dynamic processes come to equilibrium with one another.

The theory of the geomagnetic field based on the principle /30 expounded above is called the theory of the hydromagnetic dynamo (HD). The first idea of HD was proposed in 1919 by Larmor in England as an explanation for the sun's magnetism. This idea did not find any application in geophysics until the mid-'40's, when Yakov Il'ich Frenkel' in the U.S.S.R. and Walter Elsasser in the U.S.A. expressed the idea of warm thermal convection flows in the earth's core being the very cause leading to the HD process of the earth's core. Since that time, the HD theory has expanded and now it seems that the HD theory is sufficiently versatile to explain all the multitudinous phenomena associated with geomagnetism.

The electrical conductivity of the earth's interior is measured by the attenuation of geomagnetic variations which are triggered by solar activity in the upper layers of the earth's atmosphere. A varying electromagnetic signal induces in the earth varying electrical currents. As the variable current is transmitted along the conductor, a current flows in the near surface layers, in such a way that the higher the frequency, the stronger is the current "impressed" onto the surface. This phenomenon is called the skin-effect. From the theory of the skin-effect, it is possible to conclude that the depth of penetration δ of the electromagnetic variations is related to the average electrical conductivity $\bar{\sigma}$ of the layer, to the angular frequency ω, and to the velocity of light c in the medium by the following dimensional relationship: $\delta \approx c(2\pi\bar{\sigma}\omega)^{-\frac{1}{2}}$. The electromagnetic probing of the earth and the determining of its electrical conductivity $\sigma(\ell)$ as a function of the depth ℓ are based on the skin-effect theory. With the skin-effect, the lower the signal frequency, the deeper the layers which may be probed. In actual practice, the determination of $\sigma(\ell)$ encounters notable difficulties as a result of the masking influence of the oceans or of the layers of soil containing moisture and having a conductivity notably above the conductivity of rocky material out of which the earth's crust is made. The inhomogeneities or non-uniformity in the earth's crust and outer mantle also create difficulties. Nevertheless, geophysical methods have made possible the determination of the distribution $\sigma(\ell)$ down to depths of 1,000 km, at which level one must distinguish variations with a period of half a year. Scientists have succeeded in evaluating the distribution of the electrical conductivity in the lower mantle with the aid of the techniques of solid state physics and high-pressure physics. Some indications about the electrical condutivity of the lower mantle are obtained by analyzing the passage of the non-dipole part of the geomagnetic field from the core through the mantle onto the earth's surface. Our knowledge of the electrical conductivity of the mantle is summarized in Fig. 4. /31

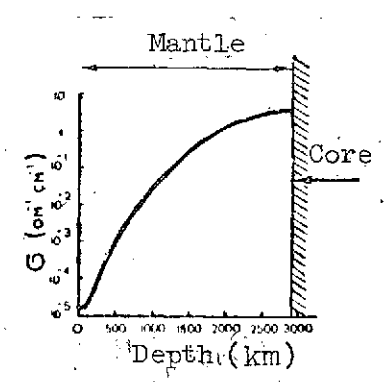

Fig. 4. The electrical conductivity of the mantle as a function of depth.

Geothermics. Temperature Distribution. Thermal Flow from the Earth's Interior

Geothermics studies the thermal state of the earth and the temperature distribution in its interior. The question of the temperature distribution is closely related to the distribution of the heat sources in the depths of the earth. Both these questions are of fundamental significance for any hypothesis about the structure and evolution of the earth. The temperature T together with the pressure p are the most important parameters for the earth's interior: the assignment of p and T determines the state of the material. Actually, many properties of the material of the earth's interior (thermal conductivity, electrical conductivity, viscosity, the dissipative function Q, the flow limit of the rocks, and other parameters) depend significantly on the temperature prevailing at a given depth. Knowledge of the temperature distribution in the earth also makes possible guidance in the choosing of one or another hypothesis of the origin of the earth. Thus, for exmaple, the hypothesis of the origin of the earth from a gas-dust cloud leads to a comparatively cold initial state for the earth, while the hypothesis of an initially molten /32
earth (the hypothesis of a hot origin) leads to significantly larger initial temperatures. As the result of the exceptionally large thermal inertia of the earth's interior, these initial differences in temperatures could not be completely obliterated during the course of the thermal evolution of the earth.

Finally, of all the observable geophysical and geological phenomena, the heat flow through the surface of the earth, from the point of view of energy, is the most significant, since the energy output associated with it per unit of time (for the whole earth it is about $6.5 \cdot 10^{27}$ ergs/year) is 10 to 100 times larger than all the energy liberated by earthquakes and volcanic activity. It is in this sense that it is said that the thermal flow from the earth's interior provides the fundamental measure of the energetics of the planet. All the remaining processes taking place in the earth's interior are, from the point of view of energy, only incidental, as it were, accompaniments to the thermal evolution of the planet.

The development of geothermics as a scientific discipline could not begin until the fundamental sources of heat in the earth's interior were discovered. Thus, the discovery of radioactivity at the end of the last century produced immediately a revolution in two geophysical disciplines: geochronology and geothermics. Actually, Lord Rayleigh already in 1906 understood the significance of radioactivity for the energetics of our planet. He made an evaluation and showed that the small traces of radioactive elements uranium and thorium (and also, as we now know, potassium) which are contained in rocks is sufficient to serve as the fundamental heat source that determines the thermics of our planet.

That the temperature of the earth's interior is high has been known for a long time. Evidence for this is given by volcanic eruptions and the temperature increase when sinking deep shafts for mining. The rate of increase in temperature with depth is called, in geophysical parlance, the geothermal gradient. In non-volcanic regions, the geothermal gradient amounts to about 3°C per 100 m of depth. The value for the geothermic gradient, generally speaking, noticeably varies from place to place, varying from less then 1°C to more than 5°C for each 100 m. On the average, at the earth's surface, the geothermal gradient amounts to 20°/km. A second geothermal quantity which can be determined experimentally is the heat flow from the earth's interior. This flow, symbolized by the letter q, is equal to the product of the coefficient κ of thermal conductivity and the temperature gradient: ∇T. /33

$$q = \kappa \nabla T \tag{12}$$

In practice, one measures the rate of temperature increase with the earth's depth ∇T and the value of κ for the rocks composing the borehole or mine shaft in which these measurements are made. Then with the aid of (12) one calculates q.

The measurement of thermal flow requires precautions since the thermal state of the outer covering for a thickness of some tens of meters is determined by meteorological factors. There are also other causes that can mask the true value of q, which characterizes the heat loss of the planet. In connection with these difficulties we have mentioned, the first precision measurements of the thermal flow on the continents were completed comparatively recently, in 1939, by Bullard in South Africa and by Benfield in England. The first measurements on the oceans are exceptionally important for geophysics since the water covering of the earth in area amounts to 3/4 of the whole surface of the planet. The measurements on the oceans gave for q about the same values that were obtained on the continents. These results were sensational and we shall speak about this later. At first, the accumulation of experimental data on the thermal flows went rather slowly. By 1960, only somewhat more than 100 measurements were known. In association with the perfection of the techniques of measurements in the ocean since the beginning of the '60's, the number of determinations of the thermal flow has begun to rise sharply. Thus, by 1965, 1,040 determinations of q were completed, and, by the middle of 1969, this number amounted to 3,560 and continues to grow at the rate of 600 determinations per year.

Averaging of the experimental data for q gives $q \gtrsim (1.2 \div 1.6) \cdot 10^{-6}$ cal/(cm^2·sec), for which the lower value is more probable. The values of the thermal flow are an integrated measure of the thermal state of the subsurface zone down to a depth of some hundreds of km. It has turned out that the different values for the thermal currents are correlated with different geological structures. Because of this, geothermal data are beginning to be used more and more widely in the physical interpretation of geological structures. /34

In contrast to the distribution of the density, pressure, and acceleration due to gravity, which are all known rather precisely, the distribution of the temperature in the earth's interior is still not precisely determined.

Evaluation of the temperature in the earth's interior is possible by means of the following concepts. The average geothermal gradient at the earth's surface is equal to 20°/km. Inasmuch as the temperature gradient does not increase with depth, then, at depths of $\ell \approx 100$ km, the temperature is not higher than 2,000°C. A more precise "thermometer" at these depths is the molten primary sources of volcanoes, for the melting points of lavas are known and are equal to about 1,200°C.

Furthermore, the mantle of the earth with respect to mechanical oscillations--seismic waves--behaves like a solid body; therefore, one can take as the upper limit of the temperatures in the earth's mantle the melting curves. Based on laboratory data, one assumes the melting point temperature at a depth of 100 km to be equal to about 1,500°C (or 1,800°K). These "reference points" make possible the determination of the distribution of the melting point temperatures in the earth's mantle, with the aid of empirical geophysical data and semi-empirical formulas for the melting curve; in particular, it is possible to estimate that at the boundary with the earth's core, the temperature of the mantle is on the order of $(5 \div 6.5) \cdot 10^3$ °K.

The earth's core is in a molten state. In view of this, it is possible to take as the lower limit of the temperatures in the core the values corresponding to the melting curve. If the core consists of iron, then, according to laboratory data, the melting point of iron at $p \approx 1.4 \cdot 10^6$ bar (this is the pressure at the mantle-core boundary) is not higher than 4,600°K. Evidently, the core does not consist of pure iron but contains an admixture of light elements, which should somewhat lower the melting point of iron. Based on these data, one calculates that the temperature at the mantle-core boundary lies in the interval of about $(4 \div 5) \cdot 10^3$ °K.

In the liquid core, the temperatures cannot be higher than the so-called adiabatic temperatures. The concept of adiabatic /35 temperatures or of adiabatic temperature gradient plays an important role in the physics of the earth and the planets. What is involved is the fact that the curve of adiabatic temperatures limits the regions for the operation of molecular and convection mechanisms of heat transport. If the temperatures are below the adiabatic temperatures (more precisely: if the temperature gradient is lower than the adiabatic gradient), then the heat transport in the medium is possible only by way of the molecular mechanism of heat conductivity. If, though, the temperatures are above the adiabatic temperatures, then convection arises--a hydrodynamic transport of fluid. The mechanism of heat transport by way of convection is a very powerful one. Therefore, if the temperatures in the earth's core were to notably exceed the adiabatic temperatures, then all the heat of the earth's core above the adiabatic heat would be quickly carried away as a result of convection to the outer boundary of the core with the earth's mantle, and the temperatures of the core would take on the adiabatic values.

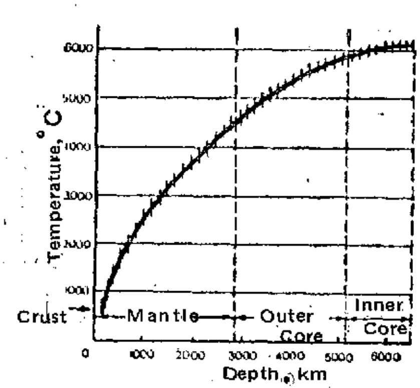

Fig. 5. Temperatures within the earth.

On the other hand, to maintain the magnetic field in the core a weak convection must be going on at all times. Consequently, the temperatures in the earth's core must be close to adiabatic. The adiabatic temperatures of the earth's core can be calculated theoretically, if only we know the temperature at the beginning of the adiabatic curve (at the mantle-core boundary). We have already said that this latter quantity amounts to about $(4 \text{ to } 5) \cdot 10^3$ °K, and this leads to a temperature at the earth's center of about $8 \cdot 10^3$ °K. The uncertainty in this value is on the order of 1,000°. The method presented above for evaluating /36 the temperature in the earth's interior may be called the "method of reference points." The "references" are the melting point of lava at a depth of 100 km and the temperature at the mantle-core boundary. The temperatures of the earth are shown in Fig. 5. The vertical strokes indicate the assumed region of uncertainty.

In conclusion, let us look at how the heat flow is formed on the continents and on the oceans. On the average, the earth's crust may be represented on the continents as a 15-km stratum of granite placed on top of a 15-km stratum of basalt. The concentration of radiogenic sources of heat in granites and basalts is well known. This permits the calculation of the heat generation in granites as $1.74 \cdot 10^{-5}$ cal/(cm^3 $\cdot$ yr) and in basalts as $0.35 \cdot 10^{-5}$ cal/(cm^3 $\cdot$ yr). The contribution from both strata to the heat flow is equal to 31 cal/(cm^2 $\cdot$ yr). If we compared this figure we have obtained with the average heat flow that annually is dissipated from the earth's surface and is equal to 30 40 cal/(cm^2 $\cdot$ yr), then we see that the heat flow is almost completely determined by the heat given off in the granite and basalt layers. Let us go now to the heat flow on the oceans. The earth's crust where there are oceans consists of a 5- to 6-km basalt stratum.The contribution to the heat flow from such a thin layer of basalt amounts to only about 2 cal/(cm^2 $\cdot$ yr).

Such a calculation had been made even before the first determinations of the heat flow on the oceans in 1956. Accordingly, it was expected that the heat flow at the oceans should be notably less than at the continents. When Bullard and, later, others obtained a value of the heat flow at the oceans coinciding with the value of the flow at the continents, this was then unexpected and was greeted with surprise in the geophysical world. The simplest explanation of this result is based on the assumption that the amount of radiogenic sources of heat per unit area both at the continents and at the seas is the same. The difference consists only in this: at the continents the sources are concentrated essentially in the outer granite and basalt strata, while at the oceans these sources are dispersed at a depth of some hundreds of km. However, this very simple explanation is not the only possible one. As a result, the question of /37 the equality of the heat flow at the continents and at the oceans remains one of the most important questions discussed in geophysics today.

RESEARCH ON GEOPHYSICAL MATERIALS UNDER HIGH PRESSURE

Direct penetration into the earth's interior is difficult. Therefore, the natural idea arises of trying to study the earth's interior by modeling it under laboratory conditions. The earth is a natural laboratory for high pressures. The pressure at the center of the earth is equal to about 3.5 million atm.,[2] and the

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[2]Let us recall the connection among the various units used for measuring pressure. 1 bar = 10^6 dyn/cm^2 = 1.01972 kg/cm^2 = 0.986324 atm. In the physics of high pressures, it is usual to use kilobars (1 kbar = 10^3 bar) and megabars (1 Mbar = 10^6 bar).

temperature reaches 6,000°C. These figures, of course, determine the range of pressures and temperatures in which one should perform modeling. Dynamic methods, using powerful shock waves for compression, cover the whole range of geophysical pressures and temperatures. With dynamic compressions, the experiment lasts fractions of microseconds, but modern equipment permits completion of all the necessary measurements. The accumulation of experimental information about the behavior of geophysical materials under high pressures and temperatures has made it possible to proceed upon a physical interpretation of the state and composition of the material of the earth's interior.

Before presenting the results of laboratory experiments, let us say a few words about geophysical materials.

Geophysical Materials

The names of the most important silicates and their mechanical parameters under normal conditions are given in Table 1.

Rocks are aggregates of minerals and it is they that we usually encounter under natural conditions. The properties of a rock are determined essentially by its rock-forming minerals. On the basis of origin, one distinguishes igneous (or "erupted"), sedimentary, and metamorphic rocks.

Sedimentary and metamorphic rocks have a secondary origin. The former are formed on the earth's surface as the result of various processes, and the latter are formed in the outer strata by means of recrystallization of other rocks. Upon the solidification of a melt inside the earth's crust, intrusive rocks are formed and, upon solidification of a melt on the surface, effusive (extrusive) rocks are formed.

These latter usually have a poorly distinguishable crystal structure or appear as amorphous bodies.

Extrusive rocks are divided according to their content of the major oxide SiO_2 into four groups: Acidic, containing 65 to 75% SiO_2; Medium (Neutral), 52 to 65%; Basic, 40 to 52%; and Ultrabasic, 35 to 40%. The average chemical composition changes in a regular fashion from the acidic to the basic rocks. In the transition to ultrabasic rocks, the contents of the oxides of light metals sharply falls and the fraction of the heavy iron-magnesium oxides increases. Accordingly, the density and elastic moduli of the rocks increase.

TABLE 1*

Table I*

Group	Name of Mineral	Formula	Density, g/cm^3	Hardness, 10^{12} dynes/cm^2	Compression modulus, 10^{12} dynes/cm^2	Velocity V_p, km/sec	Velocity V_s, km/sec	$\Phi = K/\rho$, km^2/sec^2
Quartz	Quartz	SiO_2	2,648	0,443	0,377	6.05	4,09	14.2
	Coesite		2,92	0,513	0,972	7,53	4.19	33.3
	Stishovite		4,28	1,32	3,43	11,0	5,55	80
Orthosilicates	Olivine	$(Mg, Fe)_2SiO_4$	3,3—3,5					
	Forsterite	Mg_2SiO_4	3,214	0,811	1,286	8,569	5,015	40
	Olivine	$Mg_{1,8}Fe_{0,2}SiO_4$	3,34	0,838	1.27	8,45	5,01	38
	Olivine	$Mg_{1,0}Fe_{1,0}SiO_4$	3,82	0,726	1,27	7,66	4,36	33
	Olivine	$Mg_{0,4}Fe_{1,6}SiO_4$	4,17	0,524	1,37	7,26	3,66	33
	Fayalite	Fe_2SiO_4	4,39	0,510	1,32	6,75	3,41	39
	Spinel	Fe_2SiO_4	4,85	0,814	2,05	8,05	4,10	42
	Granite	$(Ca, Mg, Fe^{+2}, Mn)_3(Al, Fe^{+3}, Cr)_2(SiO_4)_3$	3,2—4,3	0,936	1,5—1,8	8,7	4,8	40—45
Pyroxenes	Diopside	$CaMg(SiO_3)_2$	3,2—3,4	0,623	0,91—1,11	7,73	4,40	28
	Augite	$Ca(Mg, Fe, Al)[(Si, Al)O_3]_2$	3,2—3,6	0,556	0,93	7,33	4,28	29
	Jadeite	$NaAl(SiO_3)_2$	3,3—3,5		1,28			39
	Enstatite	$MgSiO_3$	3,2	0,797	1,212	8,36	4.99	38
	Hypersthene (Bronzitite)	$Mg_{0,85}Fe_{0,15}SiO_3$	3,34	0,757	1,049	7,85	4,76	31,5
	Hypersthene	$Mg_{0,7}Fe_{0,3}SiO_3$	3,44	0,725	1,073	7,70	4,59	31
	Ferrosilite	$FeSiO_3$	3,98	0,551	1,161	6,90	3,72	29

Static Investigations

The earth's crust is made up of granites and basalts, i.e., from acidic and basic rocks. The subcrustal rocks and the whole mantle of the earth consist of ultrabasic rocks. The major mineral entering into the composition of the ultrabasic rocks is olivine $(Mg, Fe)_2SiO_4$. The olivine hypothesis about the composition of the earth's mantle received wide circulation even before World War II. On the basis of this hypothesis, new important assumptions were made. In 1936 the English physicist and progressive social reformer John Bernall assumed that with increasing pressure the usual olivine rocks must experience a polymorphic transition and take on a spinel structure. In the structure of a spinel, the oxygen ions O^{-2}, as in olivine rocks, form a dense packing, not merely hexagonal, but, rather, a face-centered cubic. As a result, the density of the spinel modification increases by about 11% with respect to the density of the olivine modification. /39

*Commas in the Table are to be understood as decimal points.

The hypothesis about the olivine-spinel transition was then used by a series of geophysicists for explaining the zone of large velocity gradients for seismic waves at depths of 350 to 1,000 km in the transition zone C of the earth's mantle (see Fig. 2). For a long while, there was no success in getting the olivine-spinel transition in the laboratory. This transition was first discovered in 1958 by the Australian geochemist and geophysicist A. E. Ringwood, who obtained a spinel polymorphic modification of fayalite Fe_2SiO_4--the terminal member of the olivine series $(Mg, Fe)_2SiO_4$.

In the earth's actual ultrabasic rocks, the concentration of Mg^{+2} and Fe^{+2} ions in the olivine rocks lies in the ranges 80 to 90% and 10 to 20% respectively. However, attempts to discover an olivine-spinel transition for the magnesium-rich section or end of the $(Mg, Fe)_2SiO_4$ series were unsuccessful for a long time. Ten years of intensive effort went by in the attempt to solve this problem. During this time, broad-range discoveries were made in the investigation of silica in high-pressure apparatus. In 1953, the American physicist L. Coes synthesized the first high-density modification of quartz--coesite--under pressures equal to about 30 kbar and at a temperature of about 1,000°C. The density of coesite in a metastable state under normal conditions turned out to be equal to about 2.92 g/cm^3, that is, 0.28 g/cm^3 greater than the density of ordinary quartz. In coesite, the silicon ions Si^{+4}, as in ordinary quartz, have quadruple coordination (i.e., are four-coordinate) with the oxygen ions, and the only difference consists in the tighter packing of the ions of the silicon tetrahedron.

In 1961, the young Soviet scientists S. M. Stishov and S. V. Popova in the laboratory of academician L. F. Vereshchagin synthesized the second high-density modification of quartz. The density of this modification in the metastable state under normal conditions is equal to 4.28 g/cm^3; it has received the special name "stishovite." Stishovite was synthesized under pressures of about $1.5 \cdot 10^5$ bar and at a temperature of 1,200 to 1,400°C. In the structure of stishovite (a structure of the rutile TiO_2 type), the silicon ions are in the octahedral spaces or gaps between densely packed oxygen ions, and each ion of oxygen is /40
surrounded by three ions of silicon arrayed approximately at the apices of a right triangle. The synthesis of stishovite is one of the most outstanding achievements of geophysics during the '60's. It showed that the fundamental structural principle of the physical chemistry of silicates--the quadruple coordination of the silicon atoms with respect to oxygen--turns out to be incorrect at high pressures. At high pressures, the silicon ions Si^{+4} are in a six-fold coordination with respect to the oxygen ions O^{+2}.

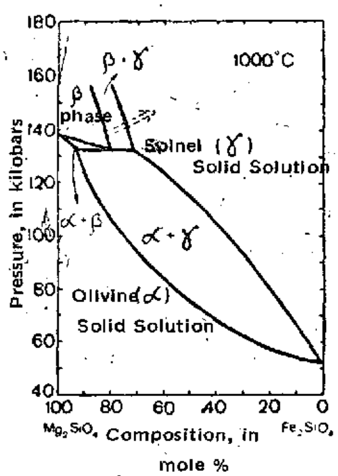

Fig. 6. Phase diagram for the system Mg_2SiO_4-Fe_2SiO_4. An isothermal cross-section is shown, corresponding to T = 1,000°C. On the abscissa is the composition in mole percent; on the ordinate is the pressure in kbar. α is the olivine phase; γ is the spinel phase; β is the modified-spinel phase. The presence of the β-phase in the diagram which we are looking at created a sensation. A remarkable feature of the Mg_2SiO_4-Fe_2SiO_4 system is the fact that the stable region for the β-phase rapidly gets wider with increasing temperature. This is shown schematically in the diagram by an arrow. The central part of the diagram is occupied by the region of the solid olivine-spinel solution ($\alpha + \gamma$). This region has the shape of a cigar. If the β-phase did not exist, then this cigar would be extended from right to left until it intersected with the pressure axis. The magnesium concentration in the actual olivines of the earth's mantle amounts to 80% and more. The temperatures at depths of about 400 km are equal to about 2,000°C. As a result, at these depths, there must occur the olivine-modified spinel ($\alpha \rightarrow \beta$) transition, and NOT the olivine-spinel ($\alpha \rightarrow \gamma$) transition that Bernall assumed. Thus, we come to a modified Bernall hypothesis.

Meanwhile, as a result of the keen competition between the Australians Ringwood and Mayor and the Japanese specialists on high pressure headed by Akimoto, the problem of studying the Mg_2SiO_4-Fe_2SiO_4 system advanced rapidly. In this competition, the Australians kept advancing somewhat ahead, outstripping their competitors literally by months, while the Japanese physicists obtained neater though more qualitative results. A complete phase diagram for the Mg_2SiO_4-Fe_2SiO_4 system was first demonstrated by Ringwood and Mayor at a symposium in Canberra (Australia) in January of 1969. This diagram, with data from Akimoto and his coworkers, is shown in Fig. 6. The detailed caption accompanying Fig. 6 explains the significance of this diagram for the problem of the internal structure of the earth. Specimens of /41 spinel (γ-phase) and modified spinel (β-phase) are capable of being kept in a metastable state under normal conditions. This makes it possible to determine the density increase in the phase transitions $\alpha \rightarrow \beta$ and $\alpha \rightarrow \gamma$. It has turned out that when $\alpha Mg_2SiO_4 \rightarrow \beta Mg_2SiO_4$ the density increases by 7.9%; and when $\alpha Mg_2SiO_4 \rightarrow \gamma Mg_2SiO_4$ the density increases by 10.8%. This latter figure was obtained by extrapolation since the γ-phase does not exist near the magnesium-end of the phase diagram (see Fig. 6).

The mineral system next in value after the olivines are the pyroxenes (Mg, Fe)SiO_3. Under normal conditions and low temperatures, the stable orthopyroxenes, which at high temperatures go over into clinopyroxenes, are rich in Mg at about 1,150°C and rich in Fe at 1,000°C. Ringwood and Mayor (in 1966) experimentally discovered the decomposition of clinoferrosilite ($FeSiO_3$) into Fe_2SiO_4 (spinel) + SiO_2 (stishovite) at a pressure of about about 100 kbar and a temperature of about 1,000°C. Akimoto and coworkers showed that the phase curve for the decomposition of clinoferrosilite is determined, in turn, by the phase curve for the transition of coesite into stishovite, which curve in the pressure (p)-temperature (T) plane is given by the straight line:

$$p\ (\text{kbar}) = 67 + 0.028\ T\ (^\circ\text{C}) \tag{13}$$

The straight line (13) separates the region of stability for the low-pressure phase (coesite) from the region of stability for the high-pressure phase (stishovite). It is of interest to note that the equation for the phase line (13), determined by Akimoto and Siono (in 1969), is very close to the first determinations of S. M. Stishov (in 1963) and I. A. Ostrovskiy (in 1965) carried out in the U.S.S.R. The olivine-spinel transition in Fe_2SiO_4 occurs in the pressure range of 45 to 55 kbar and at temperatures from 800 to 1,200°C. These transition pressures are significantly less than the transition pressures for coesite-stishovite in the same temperature range. From this, the very important conclusion was drawn that the formation of stishovite plays the determining role for the accomplishment of the reaction $FeSiO_3$ (pyroxene) → Fe_2SiO_4 (spinel) + SiO_2 (stishovite). Very recently, reports have appeared that the Japanese physicist Kavai has built a multi-stage device in which, under static conditions, pressures on the order of $2 \cdot 10^6$ bar are obtainable at temperatures of about 2,000°C. The work of Kavai is still only preliminary in nature, but, if it should be successful, it would have an extraordinarily great significance for geophysics.

Dynamic Investigations

The progress which has been made in the dynamic physics of high pressures, after World War II, is essentially due to two circumstances. First, notable achievements in the preparation of large charges of explosives were attained. As a result, experimenters currently can place blocks of explosives of various geometry and dimensions into an area of some tens of cm, at the same time preserving these dimensions with an accuracy of several microns. Second, rather accurate experimental methods (electrical /42 and optical) were worked out for recording rapid processes whose characteristic times amount to about 0.1 – 1 μsec (1 μsec = 10^{-6} sec).

Shock waves, which can be generated by powerful explosions, in passing through solid bodies create in them pressures reaching several million bar. This has substantially expanded the range of pressures for the experimental investigation of properties of solids. The most important result of this work has been the determination of the equations of state of many metals, ionic crystals, and a series of liquids and rocks up to pressures of several million bar. One of the fundamental relationships in the region of high pressures is the equation of state of a material, which determines the dependence of pressure on volume and temperature (i.e., it is a function $p = p(v,T)$). Essentially, the equation of state determines the law according to which the given material is compressed. The importance in this problem of the experimental approach arises from the fact that it is not possible to obtain for solids at present this dependence in a theoretical fashion. These investigations have a great interest for geophysics. What is involved is the fact that the pressure in the center of our planet is about $3.5 \cdot 10^6$ bar and even quite recently this pressure has turned out to be unattainable in a laboratory. Now there exists the possibility of carrying out quantitative investigations in this region of pressure and, along with this, verifying some fundamental geophysical hypotheses about the structure, composition, and state of the least-investigated portion of our planet--its core.

This sort of dynamic investigation has made possible the establishment of the equation of state $p = p(v,T)$ for iron. This made possible in 1960 the comparison of the law according to which iron is compressed with the law according to which the material of the earth's core is compressed. It has turned out that, in its properties, the material of the earth's core corresponds to the properties of iron, to within 5 to 10%, as determined from dynamic data. This result has at present led to the wide acceptance of the hypothesis of an iron core for the earth. Up until this, the hypothesis of a core made of metallized silicates had been rather wide-spread. This hypothesis was advanced by V. N. Lodochnikov in 1939 and, after the war, was developed in the work of Ramsey (the so-called Lodochnikov-Ramsey hypothesis). The essence of the hypothesis is as follows. It is known--and we have mentioned this several times--that with increasing pressure almost all materials undergo phase transitions with an abrupt density increase. On the basis of this general idea, the hypothesis was put forward that the boundary of the mantle with the core at a depth of 2,900 km is not a chemical boundary--as would take place in the hypothesis of an iron core--but a phase boundary, i.e., the silicates of the lower mantle at the boundary with the core undergo a phase transition with about a two-fold increase in density. Besides this, the hydromagnetic dynamo theory, about which we have spoken previously, requires that the material of the core possess metallic conductivity. Therefore, Ramsey assumed that during the phase transition the silicates at the same time are "metallized,", that is, they go into a metallic state. This is how the hypothesis of a core made of metallized silicates arose. In the '40's and the beginning of the '50's, the hypothesis of metallized silicates was completely invulnerable as far as its experimental vertification was concerned. However, by the beginning of the '60's, the time for the unchallenged existence of the Lodochnikov-Ramsey hypothesis came to an end.

The hypothesis was subjected to a test in the experiments completed in the U.S.S.R. by L. V. Al'tshuler and coworkers. In these experiments, shock pressure reached $5 \cdot 10^6$ bar, which is notably above the pressure at the mantle-core boundary, equal to $1.35 \cdot 10^6$ bar, and, despite this large over-pressure, not one of the rocks tried displayed a Lodochnikov-Ramsey transition. In spite of the fact that with shock waves the experiment lasts fractions of microseconds, there are still reasons for believing that the sought-after transition would be displayed were it to correspond to reality. The test of the hypothesis of a core made of metallized silicates was one of the clear demonstrations of the power of physical methods in geophysics. The use of dynamic techniques made possible the study of the properties of very important minerals and rocks under pressures and at temperatures characteristic of the D layer (lower mantle). These investigations permitted an approach to the determination of the detailed composition of the D layer. This turned out to be a complex question. What is involved is the fact that in the pressure range of 100 to 300 kbar, all silicates experience phase transitions. These phase transitions arise also with shock waves, and evidently, they notably lower the precision of the data concerning the properties of the high-pressure phase. At present, the dynamic investigations of geophysical materials under high pressures are one of the most important directions of geophysical research. /43

MODEL OF THE INTERNAL STRUCTURE OF THE EARTH

In science, when we study complex objects, we run headlong into models. People talk about models of elementary particles, models of the internal structure of stars, models of the internal structure of planets. A model is some easily visualized picture of the structure of the object under study. In constructing models, one strives to take into account everything that is known about the object considered. In step with the growth of science, models become more and more detailed, and the modern models of the internal structure of the earth are based on quite a large bulk of informative material accumulated by geophysicists up to the present. In geophysics, by a model of the earth is understood a profile of our planet in which is shown how its most important parameters vary with depth--parameters like density, pressure, gravitational acceleration, velocity of seismic waves, temperature, electrical conductivity, and others.

We have already spoken about some of these parameters. Here we shall discuss the distribution in the earth's interior of density, pressure, and gravitational acceleration. In order to understand the essence of the matter, let us begin by considering a quite simple example.

Homogeneous Model. The simplest model of our planet is the homogeneous model with $\rho = \rho(r) = \bar{\rho} = 5.52\ g/cm^3$. The value $\bar{\rho} = 5.52\ g/cm^3$ is the average density of the earth. For the homogeneous model, it is possible to calculate the distribution of the gravitational acceleration and the pressure. The gravitational acceleration g is determined with the aid of the formula known from any elementary physics course: /44

$$g = \frac{G \cdot m}{r^2} \tag{14}$$

Here $G = 6.67 \cdot 10^{-8}\ cm^3/g \cdot sec^2$ is the gravitational constant; m is the mass included within a sphere of radius r; r is the radius. In the case of a homogeneous model, the quantity m is equal to the product of the volume of the sphere of radius r times the constant value of the density $\bar{\rho}$

$$m = \frac{4\pi}{3} r^3 \bar{\varrho}.$$

Putting this into (14), we obtain:

$$g = g_0 x, \quad g_0 = \frac{4\pi}{3} G R \bar{\varrho}; \qquad x = \frac{r}{R}, \quad g_0 = 1{,}000\ cm/sec^2 \tag{15}$$

where x is the dimensionless radius, changing from 1 at the surface of the planet to 0 at the center. Consequently, in the homogeneous model, the gravitational acceleration changes linearly, decreasing from its maximum value at the surface to 0 at the center.

The pressure at a depth $\ell = (R - r)$ is equal to the weight of the rocks of the overlying layers. If the acceleration g were also constant along with the density, then the pressure at the depth ℓ would simply be equal to $\bar{\rho} g \ell$. In the general case, when the density ρ and the gravitational acceleration g depend on the depth (or, what is the same thing, on the radius), they behave in this following way. One divides the planet into arbitrarily fine spherical shells such that in each layer or shell the values for ρ and g are about constant. After thus determining the weight of rock per unit area in each layer $\rho_i g_i \Delta \ell_i$ (i is the number of the layer), one then determines the pressure by summing the weight of all the overlying layers: $p_k = \sum_{i=1}^{K} \rho_i g_i \Delta \ell_i$, where p_k is the pressure at the depth of the k-th layer (the layers are counted from top to bottom). As a result, for the homogeneous model, one obtains a quadratic dependence of the pressure on the dimensionless radius x:

$$p = p(0)[1 - x^2],$$
$$p(0) = \frac{1}{2} g_0 \bar{\rho} R = 1.73 \cdot 10^6 \text{ bar}$$

In the homogeneous model, the pressure increases in accordance with the quadratic law from zero at the surface (x = 1) to 1.73 • 10^6 bar at the center (x = 0). In the real earth, there is a notable concentration of mass towards the center (the earth has an iron core). As a result, the gravitational acceleration in the real earth diminishes notably more weakly than in the homogeneous model, and, correspondingly, the true pressure increases more sharply and at the center takes on a value about twice as large, namely about 3.6 • 10^6 bar. /45

Thus, the homogeneous model is not a very good approximation for the earth. However, for our natural satellite, the moon, the homogeneous model serves as a good approximation. We have already mentioned that, as a result of the small dimensions of the moon, the pressure at the center of the moon is small and the material at the center is compressed altogether by several percent. At the surface of the moon, the gravitational acceleration is 6 times less, i.e., g_{0m} = 16 cm/sec^2, and the pressure in a homogeneous model with the average parameters of the moon ($\bar{\rho}_m$ = 3.34 g/cm^3; R_m = 1,735 km) is p(0) = 4.71 • 10^4 bar, that is, 36.7 times less than in the homogeneous model of the earth. Consequently, the model of the internal structure of the moon can be described by the simple relationships:

$$\bar{\rho} = 3.34 \text{ g}\cdot\text{cm}^3, \quad g = g_0 x, \quad g_0 = 162 \text{ cm/sec}^2,$$

$$p = p(0)[1 - x^2], \quad p(0) = 4.71 \cdot 10^4 \text{ bar}, \quad R = 1738 \text{ km} \qquad (16)$$

Real Models (those having a distribution of density, gravitational acceleration, pressure--as opposed to assuming that some of these remain constant throughout). We shall now tell, in general features, how one builds detailed models of the internal structure of the earth, using all the geophysical information that we have. Such models are referred to, for brevity, as realistic models. The first and most substantial step on the way to building realistic models of the earth was made by the American geophysicists Adams and Williamson in 1923. They proposed to use the seismic parameter $\phi = K/\rho$ for the determination of the detailed pattern of density in the earth's interior. The seismic parameter ϕ is easy to determine from the velocities of the seismic waves v_P (1) and v_S (2), about which we have spoken in detail in the beginning of this brochure:

$$\Phi = K/\varrho = v^2_P - \frac{4}{3} v^2_S. \tag{17}$$

Since the velocities v_P and v_S are known for the earth as functions of the depth, then ϕ is also known as a function of depth. /46
The seismic parameter ϕ is equal to the ratio of the compression modulus K to the density; and, in turn, K is determined from the equation

$$K = \varrho \frac{\Delta p}{\Delta \varrho} \tag{18}$$

the ratio of the pressure increment Δp applied to the body to the corresponding density increment $\Delta\rho$ and then this ratio is to be multiplied by ρ. Thus, if we know the seismic parameter ϕ (17), then we can determine the law according to which an increment in density arises from a small increment in pressure:

$$\Delta\rho = \frac{1}{\phi} \Delta p \tag{19}$$

Now, in order to solve the problem, we must know the law by which the pressure increases in the earth's interior. The pressure increase in the earth's interior follows the hydrostatic law: namely, the pressure increment Δp attendant upon increasing the depth by $\Delta \ell$ is equal to the weight of the material of this layer pressing down per unit area:

$$\Delta p = \rho g \Delta \ell \tag{20}$$

Eliminating Δp from (19) with the help of (20), we obtain the celebrated Adams-Williamson equation:

$$\Delta \rho = \frac{\rho g}{\phi} \Delta \ell \tag{21}$$

permitting a determination of the detailed distribution of the density in the earth's interior and, correspondingly, permitting a construction or a real (or realistic) model of the earth.

At first glance, it may seem that (21) does not allow a determination of the density increment, inasmuch as here there enters the unknown function $g(\ell)$--the gravitational acceleration. Actually, $g(\ell)$ is determined by the density distribution in the planet, but this does not manifest itself in the solution of (21), since $g(\ell)$ is automatically determined when the density distribution $\rho(\ell)$ is determined.

Usually, when one speaks of the model of the earth, one first /47 has in mind the distribution of density and pressure. What is relevant here is the fact that the functions $\rho(\ell)$ and $p(\ell)$ are the starting points for the determination of many other parameters of the earth. Thus, for example, knowing $\rho(\ell)$, we can calculate the distribution of the elastic moduli in the earth ($K(\ell)$, which is the compression modulus; and $\mu(\ell)$, the shear modulus) from the velocities of the seismic waves v_p (1) and v_S (2). If $\rho(\ell)$ and $p(\ell)$ are known, then at the same time we know the equation of state for the earth's material: $p = p(\rho)$.

By comparing the $p(\rho)$ dependence determined in this way with the equation of state of various rocks and minerals as determined in laboratory experiments, we can attempt a selection of the actual material composition of the earth's interior on a quantitative basis. About ten years ago, a comparison was carried out of the function $p(\rho)$ for the earth's core with $p(\rho)$ for iron, as determined from laboratory data. The agreement of these functions to within 10% in the pressure interval of $(1.35 \text{ to } 3.6) \cdot 10^6$ bar, prevailing in the earth's core, at once was a very important indication that the central region of our planet essentially consists of iron.

In the real earth, its density does not appear to be a continuous function of the depth. From seismology, it is known that the properties of the material of the earth's interior change abruptly at the boundary of the crust and the mantle (the M boundary), and at the boundary of the mantle and core. There also exist other, though weaker, abrupt changes. Besides, in the transition layer of the mantle--zone C--the density increase arises both as a result of compression from the pressure of the overlying layers as well as as a result of the density increase in the silicate material resulting from phase transitions and their conversions to denser modifications. This latter effect is not taken into account by the Adams-Williamson equation, and, consequently, it cannot be applied to layer C. In this case it is necessary to set supplementary conditions in order to determine from them the jumps in the density at the breaks and the course of the density in zone C. Of these conditions, the most important are these two: the distribution of the density should satisfy the value of the total mass M of the earth and its average moment of inertia I*. Both these latter quantities are determined by gravimetry. Besides these fundamental conditions, one also uses /48 some others, as a result of which the density distribution in the earth at present is known with an accuracy of 1 to 2%.

At the beginning of the '20's, when Adams and Williamson proposed to use the function $\phi(\ell)$ for determining density, seismology was still in an early stage of its development. The travel or transit times of the seismic waves P and S in the earth and, correspondingly, the functions $v_P(\ell)$ and $v_S(\ell)$ themselves had at that time large errors in them. This is what forced two of the most eminent geophysicists of that time, Jeffries and Gutenberg, to resort to a reconsideration of the travel time and the distributions $v_P(\ell)$ and $v_S(\ell)$. The work lasted about ten years and was completed by the end of the '30's with new fundamental distributions for the velocities of seismic waves as determined by Jeffries and by Gutenberg. Both velocity distributions were close except for some small details.

The velocity distributions of Jeffries and Gutenberg turned out to be so accurate and good that all post-war development in seismology essentially was concerned with refining these distributions. These refinements are important for setting up a detailed structure of the mantle and core, but as far as the mechanical model of the earth is concerned, that is, its parameters $\rho(\ell)$ and $p(\ell)$, they were calculated with an accuracy of several percent by the Australian geophysicist Bullen at the end of the '30's and the beginning of the '40's. Bullen studied at Cambridge (England) with Jeffries and helped him in the quite laborious task of reviewing the transit time tables and of setting up new dependencies for $v_P(\ell)$ and $v_S(\ell)$. In 1936, when this latter task was completed, Bullen undertook the construction of new models of the earth, using the velocity distributions of Jeffries for determining the seismic parameter ϕ in the Adams-Williamson equation (21). In this construction, a fundamental role was played by the value for the moment of inertia I known at that time.

Previously we spoke in detail of how strongly the value of I affects the density distribution in the earth's interior. In

fact, Bullen, in trying out a large number of test density distributions for the earth, discovered that in order to obtain the correct value for the moment of inertia he had to introduce an anomalous density increase in zone C at depths of 400 to 1,000 km. This is now the concept of a transition layer in the earth's mantle was finally formulated. This work provided a stimulus for /49
the hypothesis on the olivine-spinel phase transitions of Bernall, which, in turn, was the starting point for the subsequent work of Ringwood. Having constructed the first modern model of the earth--the so-called Bullen model A'--Bullen introduced the concept of the division of the earth into zones, which provides a convenient frame in considering the earth's interior. The functions $\rho(\ell)$, $p(\ell)$, and $g(\ell)$ for the Bullen model A' are shown in Fig. 7. The real model of the earth presented in Fig. 7 is the culmination of the classical period in geophysics--the period of the seismology of body waves. At this period, geophysics was essentially geomechanics, inasmuch as it depended basically on the techniques developed in the mechanics of continuous media and the techniques of applied mathematics. The end of the classical period occurred by the beginning of the '50's.

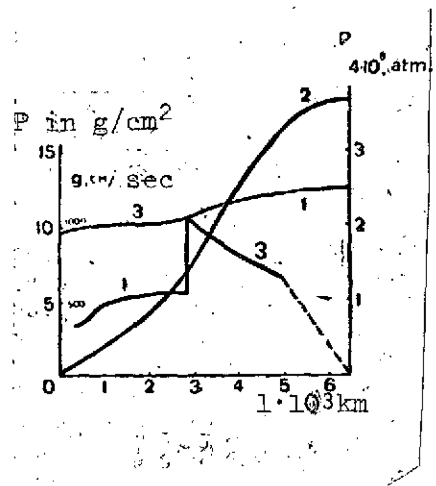

Fig. 7. The distribution of density, pressure, and gravitational acceleration inside the earth. 1--density ρ; 2--pressure p; 3--gravitational acceleration g.

The modern period in geophysics began with the work of Birch in the U.S.A. and V. A. Magnitskiy in the U.S.S.R., both of whom made the attempt to apply the techniques of solid state physics and high-pressure physics to geophysical ends. Then, Press and Ewing in the U.S.A. turned the method of surface waves into an effective means for investigating the outer layers of the earth. Then they pursued work on the natural oscillations of the earth, on the study of geophysical materials in high-pressure laborator- /50
ies, on the study of body waves with aid of seismic profiles--

profiles determined by the directions along which were placed a large number of seismographs at definite intervals. The seismic profile made possible a significantly larger sensitivity in the detection of a usable signal in comparison with single seismic receptors. And this, in turn, made possible the obtaining of a more detailed picture of the change in depth of the velocities $v_P(\ell)$ and $v_S(\ell)$.

As a result of all these innovations, the detailed structure of the upper mantle of the earth was clarified. In Fig. 8 there is shown the detailed distribution of the velocities of the transverse seismic waves $v_S(\ell)$. The fine structure of the upper mantle as shown in Fig. 8 leads to a new partitioning of the zones of the upper layers of the earth. The boundary of the outer zone--the "lithosphere" or, as it is often called, the "lithospheric layer"--is situated at a depth of about 70 km. Included in it is the earth's crust as well as the top part of the mantle. This layer combines its mechanical properties into one coherent whole. The tough lithosphere is split by about ten large blocks, along whose boundaries is arrayed an overwhelming number of earthquake foci. Beneath the tough lithosphere, in the depth interval of from 70 to 250 km is located the earth's layer of elevated fluidity. This is the earth's "asthenosphere." The viscosity of the asthenosphere is about 10^{20} to 10^{21} poise, which is a small viscosity by geophysical standards. Previously it was noted that, as a result of the small viscosity of the asthenosphere, the hard outer layers are in isostatic equilibrium: they are like gigantic icebergs floating in the earth's "asthenospheric ocean." Evidently, processes taking place in the asthenosphere determine the geological structure, of the earth's crust. Along with the flowing of materials which takes place in the asthenosphere, there are located in it the primary magma foci or sources of volcanoes. It is in the asthenosphere that there occur the processes of partial melting and formation of basaltic magma, which then is poured out onto the earth's surface via volcanic channels and cracks in the crust. Geometrically, the asthenosphere coincides with a layer of lowered seismic wave velocity in the upper mantle. This is not just happenstance, but rather has an overall cause. In the asthenosphere, the temperatures of the mantle's material approach very closely the melting point temperatures.

Beginning at a depth of about 250 km, the velocities of seismic waves start to gradually increase. This shows that, at depths of 250 to 350 km, the influence of pressure on v_S (and on v_P) prevails over the influence of temperature (from experiment, it is known that an increase in pressure calls forth an increase in the velocities v_S and v_P, while an increase in temperature leads to a decrease). At depths of 350 to 400 km (see Fig. 8), the velocity increase is anomalously large as a result of the /51

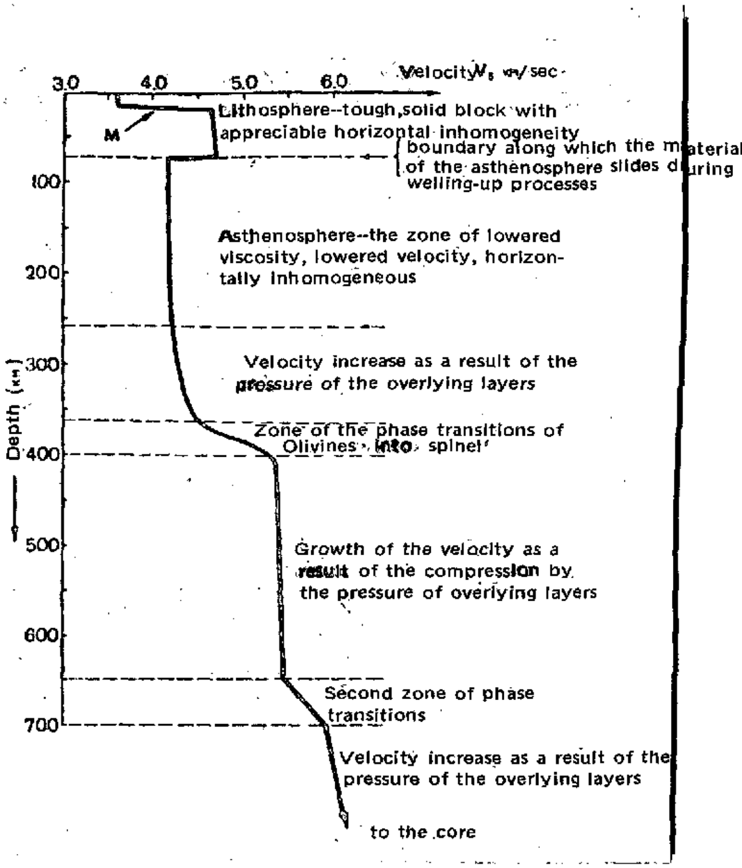

Fig. 8. Model of the earth's mantle, according to the latest data of seismology and high-pressure laboratory investigations.

phase transitions of olivines into the spinel modification--this is the first zone of phase transitions in the earth's mantle. At depths of 400 to 650 km, the velocities of the seismic waves again smoothly increase under the influence of the growing pressure of the overlying layers. At depths of 650 to 700 km (see Fig. 8), one observes the second "spurt" or sharp change in the velocities--this occurs in the second zone of phase transitions in the earth's mantle. The question is still being discussed concerning which specific phase transitions are responsible for the anomalous growth in velocity at depths of 650 to 700 km. Some believe that what is correct is the Birch-Magnitskiy hypothesis, enunciated in the early '50's, concerning the decomposition of silicates under high pressures to oxides: MgO, FeO, SiO_2 (stishovite),[3] Al_2O_3. Others, following Ringwood and Green, believe /52

[3]At the time when the oxide hypothesis was proposed to explain the anomalous velocity increase in the transition zone, the high-pressure modifications of quartz, coesite, and stishovite still had not been discovered.

that, at these depths, the fundamental rock-forming minerals of the mantle change into more complex structures. Beginning with depths of 700 km and going right to the boundary with the core, the velocities increase smoothly under the influence of the pressure of the overlying layers.

INTERNAL STRUCTURE OF THE INNER ("EARTH-GROUP") PLANETS

Mercury, Venus, and Mars belong to the inner planets. All inner planets, including the earth, are comparatively small. As a result, in the process of their formation, they could not retain the hydrogen-helium component which is the most widespread component in the cosmos. Besides this, all these planets have a deficit of water, methane, and ammonia--low-boiling and rather widely scattered compounds. The fundamental components of the inner planets seem to be silicates and iron.

Evaluation of the internal structure of the inner planets is based on geophysical data: data on the masses, radii, and moments of inertia of the planets; and data from high-pressure physics. Important ideas are adapted from modern cosmological concepts.

In spite of the scarceness of evidence about these planets, the question of their models has been treated in a series of articles and books published recently.[4]

Modern data on the mass and radius of Mercury (M_M = 0.05526 M_E; R = 2,437 km) are obtained from optical and radar observations. They give for the average density $\bar{\rho}$ the value of 5.45 ± 0.05 g/cm³. Mercury has no natural satellites; therefore there are no data about its moment of inertia. The latest radar measurements have given for the radius of the solid portion of Venus the value of 6,050 ± 5 km. Correspondingly, the average density $\bar{\rho}$ of Venus is equal to 5.25 g/cm³. The dimensions of Mars have been determined rather accurately: R = 3,380 km, and /53
its average density $\bar{\rho}$ is equal to 3.95 g/cm³. Mars has two natural satellites. This permits the determination of its dimensionless moment of inertia I* = 0.375. As a result, one obtains the following table of the fundamental parameters for the inner planets:

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[4]Our exposition is based on the work of S. V. Kozlovskiy carried out at the O. Yu. Shmidt Institute of Earth Physics of the U.S.S.R. Academy of Science.

Planet	M, g	R, км	$\bar{\rho}$, g/cm^3	I°
Mercury	$3.304 \cdot 10^{26}$	2437	5.45	—
Venus	$4.862 \cdot 10^{27}$	6050	5.25	—
Mars	$6.289 \cdot 10^{26}$	3380	3.95	0.375

One should note the difference in determining $\rho(\ell)$ for the earth and for the other planets. In the case of the earth, we know from seismology the value for $\phi = K/\rho$ as a function of the radius; and in determining $\rho(\ell)$ for the earth, we could get along without the equation of state. Moreover, with the aid of the Adams-Williamson equation, it has turned out to be possible to do the calculations for a realistic model of the earth (see Fig. 7) and thus to determine the equation of state $p = p(\rho)$ for the earth's material, using only geophysical data. In the case of the planets, the value of ϕ is not known; and therefore the equation of state $p(\rho)$ is needed, which gives the law in accord with which the material of the planet is compressed under the weight of the pressure of the overlying layers. In calculating models for Mercury, Venus, and Mars, one uses the equation of state for the earth's material and also the equation of state for Fe, MgO, FeO, Al_2O_3, and other substances--determined from dynamic and static experimental data.

MODEL OF THE INTERNAL STRUCTURE OF MERCURY

The fundamental parameters for the model of the internal structure of Mercury are presented in Table 2. Mercury must have a massive iron core, amounting in terms of mass to about 60% of the total mass of the planet. This result is quite general, inasmuch as the material in the interior of the planet is compressed comparatively weakly, about (2 to 3) $\cdot 10^5$ bar, and the existence of a concentration of iron in the planet is indicated by the large value for the average density of the planet. The question of the temperatures of the interior of Mercury is important but not yet resolved. There are substantial reasons for expecting that the temperatures in the planet's interior are high, about (2.5 to 3) $\cdot 10^3$ °K. The warming up must have arisen during the early stages of the evolution of the planet as the result of the gravitational differentiation of its initial substance into the iron core and the silicate mantle. During the time of its existence, about $4.5 \cdot 10^9$ years, Mercury could not cool off by the same process as the earth. The heat conductivity of its silicate mantle is too small to allow the internal heat outwards during cosmic intervals of time.

There is no doubt now that the time is not far off when spacecraft will set down geophysical instruments on the surface of Mercury. This will allow us to obtain new data and, correspondingly, to fill in the details in our picture of the internal structure of this planet.

TABLE 2*. MODEL OF MERCURY

Parameters of the model	Model with an iron core, g/cm^3
ρ_0 at the surface	3,29
ρ_+ at the boundary of the core	3,54—8,30
p(atm) at the boundary of core	94 580
ρ for r/R= 0.9	3,40
ρ for r/R= 0.8	3,47
ρ for r/R= 0.6	9,07
ρ for r/R= 0.4	9,49
ρ for r/R= 0.2	9,74
ρ at the center	9,8
p(atm) at the center	460 000
M_c ,%	59,8
I*	0,324
Content of metallic Fe, %	59,8

MODEL OF THE INTERNAL STRUCTURE OF VENUS

The data on Venus which can be used for the determination of /54
its internal structure are presented at the beginning of this section and are as sparse as the data on Mercury. However, Venus is a twin planet to the earth. The average radius R and the average density $\bar{\rho}$ of Venus are, in total, 5% less than the corresponding values for the earth. Therefore, the most reasonable approach in such a situation is to take, as a first approximation, the model of the internal structure of the earth for the model of the internal structure of Venus. The basic difference in the internal structure of Venus in contrast to the internal structure of the earth arises from the following consideration. Since the volume of a sphere is proportional to the cube of the radius, then with a small change δR in the radius, the relative change in the volume is equal to $3\,\frac{\delta R}{R}$. In changing from the earth to Venus, the relative decrease in radius amounts to $\frac{\delta R}{R} \sim 5\%$ and

*Commas in the Table are to be understood as decimal points.

the corresponding decrease in volume $\frac{\delta V}{V} = 3\,\frac{\delta R}{R} \sim 15\%$. The average density $\overline{\rho}$ is determined as equal to the ratio of the total mass M to the total volume of the planet. The average density of Venus is, in total, 5% less than the average density of the earth, but this is small in comparing the two planets since the volume of Venus is less than the volume of the earth by 15%. From this comes an important conclusion: the material of the Venusian interior must be somewhat heavier than the material of the earth's interior; i.e., it contains more iron. /55

There are three possible reasons contributing to the weight of Venus: (A) the iron core of Venus could be about 5 to 10% larger relative to that of the earth; (B) the density of the silicates in the mantle could be 5 to 10% greater than for the earth; i.e., in them the Fe/Mg ratio is higher; and (C) the third possibility is a combination of these two: the core of Venus is larger <u>and</u> the density of the silicate material of the mantle is somewhat higher than for the earth. To make a choice among these possibilities, one needs supplementary data. One may hope that in the '70's an artificial satellite of Venus will be launched. Then from the data on the gravitational field of the planet one will succeed in determining its dimensionless moment of inertia I*, and this will immediately allow a determination of the specific zone of the Venusian interior where the greater weight of the material of the planet is being caused in comparison with the material of the earth.

MODEL OF THE INTERNAL STRUCTURE OF MARS

For Mars, the value of the dimensionless moment of inertia I* is known, and this is quite significant for building its models. Kozlovskaya has considered two choices for models of Mars for different thicknesses of the crust: 1) Mars-1, with a minimal crustal thickness (20 km), whose mass amounts to 1.3% M. Such a crust for Mars is equivalent to an earth crust 18 km thick if one assumes that the earth's crust is distinct from the 1,000-km-thick upper stratum of the mantle and that the crust of Mars is distinct from all the strata of the mantle; 2) Mars-2, with a maximal crustal thickness (216 km), whose mass amounts to 13% M. Such a crust corresponds to the assumption that there has taken place a melting of all the silicon-aluminum material from all the strata of Mars. Calculations have been carried out based on the equation of state for the earth's mass. To satisfy the value of I* = 0.375 known for Mars, it is necessary in the Mars-1 model to make denser the reference material of the earth's mantle by 6% and the iron core and its mass by 6.3%.

TABLE 3*

Density distribution in the model of Mars

Mars-1			Mars-2		
Reference model: Earth-1, $\rho_{vpl}=1.061\rho_{ref}$			Reference model: Earth-1, $\rho_{vpl}=1.137\rho_{ref}$		
r/R	$M(r)/M$	ρ, г/см³	r/R	$M(r)/M$	ρ, г/см³
Crust			Crust		
1,000	1,000	2,8	1,000	1,000	2,78
0,994	0,987	2,8	0,936	0,873	2,85
Mantle			Mantle		
0,994	0,987	3,515	0,936	0,873	3,841
0,977	0,943	3,526	0,864	0,701	3,921
0,909	0,778	3,618	0,792	0,551	4,014
0,854	0,661	3,674	0,740	0,458	4,074
0,765	0,494	3,765	0,673	0,353	4,140
0,661	0,341	3,858	0,585	0,240	4,397
0,565	0,232	4,105	0,465	0,126	4,658
0,442	0,131	4,355	0,335	0,0500	4,843
0,369	0,0902	4,476	0,243	0,0212	4,928
0,330	0,0738	4,529	0,206	0,0142	4,957
0,297	0,0627	4,575	0,141	0,00698	4,997
Iron Core			Iron Core		
0,297	0,0627	9,326	0,141	0,00698	9,397
0,276	0,0506	9,371	0,125	0,00504	9,410
0,253	0,0391	9,417	0,113	0,00385	9,422
0,226	0,0282	9,467	0,100	0,00281	9,434
0,194	0,0181	9,518	0,0863	0,00193	9,446
0,176	0,0135	9,549	0,0712	0,00125	9,459
0,155	0,00927	9,580	0,0634	0,00100	9,465
0,130	0,00559	9,611	0,0558	0,000806	9,471
0,0992	0,00263	9,642	0,0364	0,000505	9,489
0,0593	0,000755	9,673	0,0278	0,000441	9,502
0	0	9,692	0	0	9,540

Thickness of crust = 20 km
Radius of core R_c = 1,000 km

P_{atm}
Base of crust 2,470
Boundary of core $2.73 \cdot 10^5$
Center $4.05 \cdot 10^5$

Thickness of crust = 216 km
Radius of core R_c = 475 km

P_{atm}
Base of crust 22,176
Boundary of core $3.00 \cdot 10^5$
Center $3.89 \cdot 10^5$

*Commas in the Table are to be understood as decimal points.

For the Mars-2 model, one needs to make denser the silicate material of the earth's mantle, which is taken as our reference, by 13.7%. The metallic iron core in this model is very small and amounts to less than 1% by mass (M< 1%M). The density increase in the Fe/Mg ratio in the Martian silicates in comparison with those of the earth. The Mars-1 and Mars-2 models should be viewed as limiting models of the planet. Their parameters are presented in Table 3. The real model of Mars possibly occupies an intermediate position between the Mars-1 and Mars-2 models. To progress further in this question, we need seismic data about the interior of Mars. Evidently, such data will be obtained in the '70's.

INTERNAL STRUCTURE OF THE GIANT PLANETS JUPITER AND SATURN

Jupiter and Saturn are hydrogen-helium planets. To be convinced of this, it is sufficient to turn to Fig. 9, in which are presented the mass-radius curves for planets consisting of pure hydrogen and pure helium. We see that Jupiter as well as Saturn are quite close to the hydrogen curve. This circumstance, of course, is not accidental. Hydrogen is the most widely distributed element in the solar system, in the stars, and in interstellar space; and the gravitational field of the giant planets is such that it is capable of retaining a hydrogen atmosphere during the course of existence of the planets. /57

The second most abundant element in the universe is helium. The abundance of helium, by number of particles, is such that the ratio H/He $\approx$ 10. Turning to Fig. 9, we see that the planets Jupiter and Saturn are somewhat shifted relative to the hydrogen curve towards the helium curve. In this regard, it is natural to expect in both planets the presence of an admixture of helium. The determination of the helium concentration in both planets is a very important task of the physics of planets and has a great significance in cosmology. This sought-after ratio is established if a model of the planet is constructed. The abundance of the remaining elements in the solar system--for example, oxygen, carbon, nitrogen, silicon, iron, etc.--is quite small in comparison with the abundance of hydrogen and helium, and their contents in Jupiter and Saturn are currently determined with less certainty than the contents of hydrogen and helium.

The building of planet models is based on astronomical data on the planets and on data from high-pressure physics on the equations of state, especially the equations of state of hydrogen and helium. The latter were obtained by DeMarcus in the U.S.A. and V. P. Trubinsyn in the U.S.S.R. Previously, we discussed in detail the determination of the earth's gravitational field with the aid of artificial satellites. Jupiter and Saturn have ample natural satellite systems. Thus, Jupiter has 12 natural satellites and Saturn has 10. Observation of the motion of the satellites nearest to these planets has permitted determining the /58
first two gravitational moments J and K in the gravitational potential of Jupiter and Saturn:

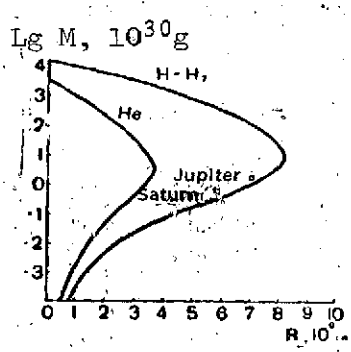

Fig. 9. Mass-radius diagram from planets consisting of only hydrogen and only helium.

$$V(r, \Theta) = -\frac{GM}{r}\left[1 - \frac{2}{3} J \left(\frac{a}{r}\right)^2 P_2(\cos\Theta) + \frac{4}{15} K \left(\frac{a}{r}\right)^4 P_4(\cos\Theta) + \ldots\right]$$

Here, G is the gravitational constant; r is the distance from the center of the planet; θ is the polar distance; $P_2(\cos\theta)$ and $P_4(\cos\theta)$ are the second and fourth Legendre polynomials (about which we have already spoken in detail in this brochure).

The formulas in the theory of the shape of liquid rotating planets combine the distribution of certain physical parameters (density, pressure) in the planet's interior with values of the average density and the multipole moments J and K (determined from observations). The task is as follows: to find that distribution of density which best satisfies the observational data.

The data for the distribution of the pressure, density, and mass in Jupiter and Saturn are presented in Table 4. These data are presented for values corresponding to a relative radius (r/R) from 1 down to 0.2. Below this--i.e. at the center region of the planets--the distribution of ρ and p are insufficiently determined.

TABLE 4*. MODELS OF THE PLANETS JUPITER AND SATURN

Relative radius, r/R	Jupiter: Pressure p, Mbar	Jupiter: Density ρ, g/cm³	Jupiter: Relative mass, M(r)/M	Saturn: Pressure p, Mbar	Saturn: Density ρ, g/cm³	Saturn: Relative mass, M(r)/M
1,0	$3{,}0 \cdot 10^{-6}$	$5{,}5 \cdot 10^{-4}$	1,0	$3{,}0 \cdot 10^{-6}$	$5{,}5 \cdot 10^{-4}$	1,0
0,995	$3{,}7 \cdot 10^{-4}$	0,0164	0,99995	$3{,}4 \cdot 10^{-5}$	$3{,}0 \cdot 10^{-3}$	0,99998
0,99	$2{,}8 \cdot 10^{-3}$	0,055	0,9996	$2{,}2 \cdot 10^{-4}$	0,0114	0,9998
0,98	0,020	0,147	0,997	$1{,}44 \cdot 10^{-3}$	0,039	0,9980
0,96	0,095	0,28	0,988	$7{,}9 \cdot 10^{-3}$	0,092	0,995
0,94	0,23	0,40	0,973	0,028	0,169	0,983
0,92	0,38	0,47	0,957	0,055	0,22	0,970
0,9	0,57	0,55	0,941	0,088	0,27	0,953
0,85	1,29	0,76	0,878	0,21	0,38	0,900
0,8	2,2	0,96	0,815	0,37	0,47	0,841
0,75	3,5	1,39	0,74	0,58	0,56	0,78
0,7	5,2	1,63	0,64	0,86	0,65	0,70
0,65	7,2	1,84	0,55	1,16	0,73	0,63
0,6	9,5	2,1	0,46	1,53	0,81	0,58
0,55	11,9	2,3	0,38	2,0	0,92	0,51
0,5	15,0	2,5	0,30	2,5	1,24	0,46
0,4	20,5	2,9	0,19	4,2	1,49	0,33
0,3	26	3,3	0,11	6,9	1,81	0,26
0,2	34	3,7	0,05	11,5	2,2	0,20

*Commas in the Table are to be understood as decimal points.

Under ordinary conditions, hydrogen is a diatomic gas H_2. /59 If this gas is compressed, then its density will increase and it is turned into a liquid, and, at still higher pressures, it can solidify if its temperature becomes lower than the melting point at some pressure. However, at pressures less than $2.5 \cdot 10^6$ bar, the liquid and solid hydrogen, as is the case for gaseous hydrogen, consist of H_2 molecules. Therefore, one says that hydrogen at $p \lesssim 2.5 \cdot 10^6$ bar is in the molecular phase. At the pressure of $2.5 \cdot 10^6$ bar, the molecular hydrogen experiences a phase transition and is converted into metallic hydrogen. In this transition, the molecular bond is destroyed and the metallic hydrogen is a monovalent metal, whose nucleus consists of a single proton, and the electrons--as is usual in metals--are delocalized and form a conducting electron liquid.

By its properties, metallic hydrogen should be similar to alkali metals, such as Li, Na, and K; and it is possible to say that metallic hydrogen is the simplest alkali metal. The density of metallic hydrogen at $2.5 \cdot 10^6$ bar is close to 1 g/cm^3. If we turn to Table 4, we see that the outer mantle of Jupiter with a thickness of about 15,000 km is in the molecular phase; while the remaining part of the planet consists of metallic hydrogen. In Saturn, the molecular mantle extends down to half the radius; below this, there is a mantle consisting of metallic hydrogen down to $r/R \approx 0.2$; and, finally, there is the central core of the planet consisting of other chemical substances from outer space (H_2O, CH_4, NH_3, MgO, FeO, and others).

Calculations show that in the hydrogen-helium mantle of both planets, outside of the heavy central core, the concentrations, by mass, of hydrogen X, helium Y, and other volatile gases (H_2O, CH_4, NH_3) Z amount to

$$X = 0.8 \quad Y = 0.18 \quad Z = 0.02,$$

and, along with this, the core of Jupiter amounts to 2% by mass and the core for Saturn amounts to 20%. Such a small core for Jupiter is quite indefinite and lies within the limits of uncertainty of the calculations. The interior of Jupiter and Saturn must be presumably strongly heated. The best estimated temperatures for these planets are the adiabatic temperatures. These adiabatic temperatures, as well as the melting point temperatures, are easy to calculate if we know the distributions of the density and of the pressure. It has turned out that in the hydrogen-helium planets the adiabatic temperatures are notably higher than the melting temperatures and this means that both Jupiter and Saturn are in the gas-liquid state--the material in their interiors is at temperatures higher than the corresponding melting points. The temperatures in the centers of both planets are about 20,000°K.

INTERNAL STRUCTURE OF THE GIANT PLANETS URANUS AND NEPTUNE

The giant planets Uranus and Neptune occupy an intermediate position between the planets of the inner group--made of silicates and iron--and the hydrogen-helium planets Jupiter and Saturn. The most important characteristic of these giant planets is that they consist essentially of hydrogen compounds (water, methane, ammonia, and others) while silicates, iron, hydrogen, and helium are contained in them only in small quantities. In this connection, the building of the models of Uranus and Neptune /60 is the most complex of the problems in the physics of planets.

The difficulties can be divided into four groups: 1) the unreliability of the reference astronomical data; 2) lack of certainty about the primary composition; 3) uncertainty in the equations of state; and 4) the large number of components from which the planets are made up. All this makes the building of real models hard, and these very models for these planets are still quite crude.

Uranus and Neptune are at the periphery of the solar system, and their properties are quite essential for making judgments about the development of the periphery of the gas-dust cloud from which the planetary system was formed. Let us now consider that primary material composition from which, presumably, Uranus and Neptune have been formed. Both planets were formed at the periphery of the gas-dust cloud, where the temperatures always were low and the primary composition did not undergo any substantial chemical transformations. According to data from space chemistry, silicon and magnesium form oxides at the low temperatures at the periphery of the planetary system, while carbon, nitrogen, and sulfur form hydrides. Iron and nickel remain in the uncombined state, although this question is still not completely clarified, and, possibly, iron and nickel also may form oxides. However, since the quantity of iron and nickel is small, it is not essential in calculating models to take into account the iron-nickel component in any explicit form.

In Table 5 is presented the relative distribution of various compounds. In Table 6 are presented the corresponding melting points. Uranus is 19 times more distant from the sun that the earth; and Neptune is 30 times more distant. The temperature of the gas-dust protoplanetary cloud in this distant zone of the solar system must amount to, according to calculations, only 50°K. According to data on melting points, as given in Table 6, at temperatures of 50°K, hydrogen, helium, and neon will be in the gas phase (in a vacuum--i.e., at zero pressure), while the remaining materials will be in the solid state (in the form of ices).

TABLE 5*. RELATIVE ABUNDANCE OF COMPOUNDS

Element or Compound	Relative Mass Abundance
H	1004
He	400
Ne	10,1
H_2O	14,1
CH_4	6,39
NH_3	1,91
H_2S	0,760
Ar	0,303
SiO_2	1,90
MgO	1,01
Fe + Ni	0,260

TABLE 6*. MELTING POINT TEMPERATURES

Element or Compound	Melting Point Temperature, °K
He	Liquid at 0°K
H_2	14,1
He	24,5
Ar	84
CH_4	89
NH_3	196
H_2S	190
H_2O	273

*Commas in the Table are to be understood as decimal points.

Accordingly, all materials enumerated in Table 5 are divided into three groups: 1) gaseous component labelled HHeNe (i.e., consisting of H_2, He, Ne); 2) ice component, consisting of H_2O, CH_4, NH_3, H_2S, and Ar; 3) a component consisting of the oxides of magnesium, silicon, and of iron-nickel--denoted in short as the HC or heavy component. Moreover, with the aid of experimental data that we have, one constructs the equations of state, that is, the function $p(\rho)$, both for the ice component and for the HC component. For the gaseous component, consisting essentially of hydrogen and helium, the function $p(\rho)$ is known and we have already spoken of this in a previous section. Models of these planets are constructed from the three groups of materials which we have listed in such a way that the models satisfy the average density of the planet and--in the case of Neptune--the known gravitational moment J. As a result, two-layer models for Uranus and Neptune have been constructed. In these models, the central core consists of a mixture of ice and HC components. This mixture is for brevity indicated as the HCI components. However, the outer layer or mantle is gaseous, made up of the hydrogen-helium components. The distribution of density and pressure in the models of Uranus and Neptune is given in Table 7. The pressure in the centers of Uranus and Neptune is equal to $5.9 \cdot 10^6$ bar and $7.15 \cdot 10^6$ bar, respectively. The transition from the gaseous covering to the HCL component in Uranus occurs at $r/R = 0.733$ and in Neptune at $r/R = 0.757$ ($R_U = 25{,}200$ km, $R_N = 25{,}000$ km). /61

The models of the planets, presented in Table 7, are also characterized by the following parameters. In Uranus, the gaseous covering amounts to 10.4% of the total mass of the planet, and in Neptune, 11.7% of the total mass. The total content of hydrogen (both free and combined) amounts to 19.6% of the total mass for Uranus and 20% of the total mass for Neptune; while as for water in both planets, it amounts to about half the total mass. Thus, the construction of models for the planets has led to the conclusion--unexpected at first glance--that both Uranus and Neptune consist essentially of water. The best estimated temperatures of the giant planets are the adiabatic temperatures. Calculation of the adiabatic temperatures for Uranus and Neptune gives a temperature in their centers of about $(2.5 \text{ to } 3) \cdot 10^4$ °K. It has turned out that the adiabatic temperatures of these planets over their whole range exceed the calculated melting point temperatures. From this, the important conclusion was made that both planets are in the gas-liquid state. Their outer coverings or /62 mantles consist of essentially gaseous-liquid hydrogen in the molecular phase, while the central core consists of melted ice components.

TABLE 7*. MODEL OF URANUS AND NEPTUNE

r/R	Uranus		Neptune	
	Pressure p, Mbar	Density ρ, g/cm^3	Pressure p, Mbar	Density ρ, g/cm^3
0	5,90	4,36	7,15	4,52
0,10	5,76	4,32	6,97	4,52
0,20	5,33	4,22	6,44	4,48
0,30	4,65	4,06	5,58	4,28
0,40	3,78	3,81	4,50	3,99
0,50	2,61	3,48	3,27	3,64
	2,29	3,29	2,63	3,42
0,60	1,76	3,07	2,00	3,17
0,65	1,23	2,82	1,39	2,90
0,70	0,776	2,53	0,824	2,57
0,757			0,28	1,97
0,757			0,28	0,447
0,773	0,207	1,83		
	0,207	0,413		
0,80	0,168	0,390	0,206	0,412
0,85	0,109	0,352	0,135	0,365
0,90	0,061	0,310	0,0775	0,327
0,95	0,023	0,255	0,031	0,274
1,00	0	0,105	0	0,105

INTERNAL STRUCTURE OF THE MOON

It was noted previously that the homogeneous model was a good approximation for the moon. Thus, the distribution of pressure in the interior of the moon is given by formula (16). As present, the first probings of the moon are being carried out with geophysical techniques. Let us summarize, in brief, the results of this work.

With the help of three seismometers placed on the moon by Apollos XII, XIV, and XV, it has been possible to determine the velocity profiles of the longitudinal seismic waves v_p up to depths equal to 80 km. One should consider as the most interesting property of the seismic model of the moon's outer stratum the presence of a stratified crust to a thickness (or width) of 65 km. Measurements have been carried out in the region of the Fra Mauro crater in the Ocean of Storms. The seismic cross-section is characterized by the following details:

*Commas in the Table are to be understood as decimal points.

a rapid increase in the velocity v_p from a value of 0.1 km/sec on the surface to 5 km/sec at a depth of 10 km; 2) a sharp increase in the velocity at a depth of about 25 km (boundary of a division in the moon's crust); 3) approximately constant value of $v_p \sim 7$ km/sec between depths of 25 km and 65 km; 4) abrupt increase in velocity at the foundation of the moon's crust (depth of 65 km); 5) in the subcrustal region--what would be the moon's mantle region-- the velocities are quite high, exceeding 9 km/sec.[5] /63

The interpretation of the velocity profile is as follows. Close to the surface, to depths of 1 to 2 km, the velocity behavior corresponds to a stratum of dust condensed under the action of its own weight--a stratum formed from fragmentation of rock-- breccia and rock debris. Below 2 km, down to 25 km, the stratum of the moon's crust should consist of basalts of the same type as was brought to the earth during the flights of Apollo XI and Apollo XII. This conclusion is based on experimental data on the velocities of longitudinal waves in the moon's basalt. The large gradients for the velocity increase at depths to 10 km is brought about by the effect of the pressure on the rocks containing cracks and pores.

The second stratum of the moon's crust should differ remarkably from the specimens of lunar basalt brought from the surface of the moon to the earth. The terrestrial equivalent of these rocks are rocks of the gabbro or pyroxine or anorthosite types.

The velocities in the mantle, determined still only qualitatively, are notably larger than the velocities in ordinary earth rocks but are close to those in olivines enriched with magnesium.

A moonquake with a focus at a depth of 800 km has been recorded. This shows that, at least down to this depth, the interior of the moon is in the solid state.

With the aid of two instruments, the heat flow from the interior of the moon has been determined. It has turned out to be equal to $0.79 \cdot 10^{-6}$ cal/cm^2 $\cdot$ sec, which amounts to about half of the heat flow from the earth's interior. It is interesting to note that this result practically coincides with the result obtained by radio techniques in the pre-sputnik era by V. S. Troitskiy: $q \sim (0.85 \pm 0.2) \cdot 10^{-6}$ cal/cm^2 $\cdot$ sec. The result of V. S. Troitskiy in its time (about ten years ago) was subject to doubt and criticism since it exceeded by a factor of 2 to 3 the heat flow in the chondrite model of the moon. Nonetheless,

[5] Let us recall that in the cover of the earth's mantle, the velocities are equal to about 8 km/sec.

the result of V. S. Troitskiy was verified by direct measurements on the moon, and this result places in doubt the hypothesis of the chondrite composition of the moon.[6]

The first electromagnetic probes of the moon have been made. These probes have made possible an evaluation of the electrical conductivity, in its crude details, of the moon's interior. On the basis of electrical conductivity, the moon's interior can be divided into three zones: 1) an outer non-conducting stratum to a depth of about 70 km--the crust; 2) an intermediate stratum--the mantle from 70 km down to a depth of about 700 km has an average electrical conductivity $\sigma_m \sim 10^{-4}$ ohm^{-1} • meter^{-1}; 3) the central zone--the core of the moon--has an average electrical conductivity $\sigma_{av} \sim 10^{-2}$ ohm^{-1} • m^{-1}. For the olivine model of the moon, one may add to these estimates of the electrical conductivities the temperatures: for the crust, < 440°K; for the mantle, 840°K; and for the core, about 1,240°K.

CONCLUSION

The start of the conquest of space introduced new problems in practically all realms of human knowledge. The launching of the first artifical satellites forced us to look quite differently even at geophysics. It became quite apparent that that path which geophysics traversed in the study of the earth would be extended in outer space for studying other planets. This new path will be more arduous and more lengthy but it is unavoidable. According to the hypothesis of our correspondent academician Otto Yul'evich Shmidt, the planetary system was formed in a single process of evolution from a pre-planetary gas-dust cloud. The moon and planets, strictly speaking, as well as meteors, are those silent companions of our native planet which definitely help establish for geophysicists the most important features of evolution of the earth. For the first time, we are trying as far as possible to "lift up" the problem of the internal structure of planets to the modern level of knowledge. In this, one uses geophysical experiments; construction of models of the earth; evidence about the distribution of the elements which we take from astrophysics and the study of meteorites; data on the masses and dimensions of planets obtained from astronomical observations; and also data on the behavior of materials at high pressures and temperatures. /64

[6] Chondritic meteorites, or chondrites, are--according to this hypothesis--the primary undifferentiated material out of which the inner planets and the moon are made.

It is natural with the increase in the development of our space program for investigating the moon and planets that we shall be better able to work out detailed theories of their internal structure and at the same time check and verify our current concepts. In the meanwhile, we tend to agree with the first models of the inner planets (Mercury, Venus, Mars) and have constructed models of the giant planets (Jupiter, Saturn, Uranus, Neptune). In this, we have not avoided some unexpected results. Thus, it has turned out that all the giant planets are liquid and their interiors are quite hot. This is tied in with the high compressibility of the hydrogen-helium giants Jupiter and Saturn and of the planets Uranus and Neptune, which are one-half water in composition. Especially surprising has been the fact that very high temperatures, exceeding 20,000°, must be reached in the interior of Uranus and Neptune--planets situated at the periphery of the solar system.

When this brochure was completed, the author received a letter from Doctor D. V. Dunham, a coworker at the MacDonald Observatory of the University of Texas in the U.S.A. In this letter, it was reported that, having analyzed a long series of observations of a natural satellite of Uranus (Ariel), he had succeeded in determining the gravitational moment of Uranus, J_2, which gives the first correction term to the Newtonian gravitational field of the planet. This J_2 obtained from observations has turned out to be in excellent agreement with the value for this quantity obtained purely theoretically in model computations carried out in the Institute of Earth Physics of the U.S.S.R. Academy of Sciences and published in 1971.

This communication is a highly promising start for the scientific cooperation of theoreticians and experimentalists occupied with the study of the outer planets of the solar system. The specific value of J_2 determined by Dunham is such that it points to a notable concentration of material towards the center of the planet, that is, Uranus--like the earth--has a core. This core, as was shown in a basic part of this brochure, must consist essentially of water.